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1 **Distinct sources for high-K and adakitic magmatism in SE Iran**

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12

13 **Abstract**

14

15 *Research into Arabia-Eurasia collision zone magmatism in Kerman Province, SE Iran, has largely*
16 *focused on Late Cenozoic adakitic stocks or domes, with debate around lower crustal or subducted*
17 *slab origins. Contemporary hawaiite-trachyandesite lava flows have been overlooked. New analyses*
18 *for domes and lavas from near Dehaj show major and trace element distributions relating to two*
19 *distinct compositional series. One contains medium-K domes with SiO₂ > 60 wt.%, high Sr/Y and*
20 *La/Yb and generally low MgO, Ni and Cr, showing high-silica adakite affinity. The other series has*
21 *high-K affinity and includes both lavas and dome samples. The two suites partially mixed in the*
22 *shallow crust, confirmed by fieldwork and petrography. Isotopically the two suites are*
23 *indistinguishable, implying a geologically ‘young’ age for the source of the adakites. Given its*
24 *geochemical signatures and non-relationship with the largely mafic, mantle-derived high-K series, we*

consider the adakite series to be derived from melting of eclogitized mafic lower crust. The high-K series relates to dehydration melting of mantle peridotite deeper within the ~220 km thick lithosphere. We also explore adakitic magmas across Iran and their relationship to porphyry copper deposits. At Dehaj and several other Iranian centres, adakites are chemically controlled by garnet as a source or fractionating phase, and are barren, whereas the presence of amphibole as a key phase seems to correlate with Cu mineralisation. This study also shows the need for evidence from multiple datasets to constrain adakite genesis and warns of avoiding sampling bias towards felsic lithologies.

Keywords

Iran, adakite, petrogenesis, collision, magma mixing; mineralisation

1. Introduction

Late Cenozoic magmatism is commonplace across much of Iran, Turkey and the Caucasus (Fig. 1) and results from a variety of geodynamic and petrological processes including Neo-Tethyan slab break-off (Keskin, 2003; Zor et al., 2008), localised lithospheric detachment and sub-lithospheric convection (Kaislaniemi et al., 2014), crustal extension related to strike-slip faults (Sadaat et al., 2010; Shabanian et al., 2012) plus intra-lithospheric breakdown of hydrated peridotite (Allen et al., 2013; Sugden et al., 2019). The diversity of mafic to felsic magmatic products formed in the past few million years is high, ranging from alkaline to sub-alkaline, of within-plate to arc-like chemistry, and moderately to highly enriched incompatible trace element concentrations (Pearce et al. 1990; Saadat and Stern, 2012; Pang et al., 2012, 2013; Kheirkhah et al., 2013, 2015; Allen et al., 2013; Sugden et al. 2019). Many publications on Late Cenozoic samples identify trace element characteristics consistent

with small volume melting of lithospheric mantle previously modified during subduction of the Neo-Tethyan oceanic slab(s) (Pearce et al., 1990; Keskin et al., 1998; Özdemir et al., 2006, 2014; Neill et al., 2015). The extent to which magmas interact with the diverse crustal lithologies also varies considerably but can be reasonably detected in Iran using isotopic data: large parts of the predicted lower crust are ultimately Gondwanan in origin and are sufficiently old to present high $^{87}\text{Sr}/^{86}\text{Sr}$ and low $^{143}\text{Nd}/^{144}\text{Nd}$ ratios compared to mantle-derived basalts (Davidson et al., 2004; Ghalamghash et al., 2016; Pang et al., 2012). Such magmatic diversity may also be driven by the region's strongly heterogeneous lithospheric thicknesses, ranging from lithosphere over 200 km thick in the heart of the Zagros core, to less than 50 km in parts of Eastern Anatolia, where mantle lithosphere is very thin or absent (Priestley et al., 2012). Therefore, Iran is a good test site for the identification of sources and processes involved in collision magmatism. It is also important to understand magma genesis in a region where magmatic activity controls mineral resources (Zarasvandi et al., 2015), natural hazards (Karakhanian et al., 2002; Mortazavi et al., 2009), and geothermal energy sources (Seyedrahimi-Niaraq et al., 2017).

One group of rocks which has received special attention in Iran is the adakite series. These volcanic or intrusive lithologies have attracted much controversy over their definition, petrogenesis, and implications for the occurrence of porphyry Cu mineralisation (e.g. Jahangiri, 2007; Omrani et al., 2008; Aftabi and Atapour, 2009; Khodami et al., 2009; Shafiei et al., 2009; Richards et al., 2012; Asadi et al., 2014; Azizi et al., 2014; Zarasvandi et al., 2015; Pang et al., 2016; Shaker Ardakani, 2016; Lechmann et al., 2018). Omrani et al. (2008), Pang et al. (2016) and Shaker Ardakani (2016) described Late Cenozoic adakite-like felsic rocks in Kerman Province and elsewhere in Iran (Fig. 1) and argued that these magmas formed due to melting of the Neo-Tethyan slab beneath the collision zone (Omrani et al., 2008), or by melting of existing arc rocks in the lower crust (Pang et al., 2016). We present new

geochemical data from around Dehaj in Kerman Province which are used to a) debate the origins of the adakitic signature, b) link the origin of the adakites to what is known about contemporaneous non-adakitic magmatism and the crustal and lithospheric thickness in Kerman Province and c) identify any geochemical patterns across the regional distribution of adakitic magmatism and its correlation with porphyry copper mineralisation.

2. Geological background

2.1. Regional geological setting

Kerman Province is part of the Turkish-Iranian Plateau, the latter an area of $\sim 1.5 \text{ M km}^2$ and $\sim 1.5\text{-}2 \text{ km}$ elevation within the Arabia-Eurasia collision zone (Agard et al., 2011). The plateau was created by the collision and postdates the extensive marine limestones of the Late Oligocene - Early Miocene Qom Formation and lateral equivalents (Reuter et al., 2009). Plateau elevations are partly an isostatic response to crustal thickening, but other mechanisms may include the influence of warm and low-density upper mantle (Maggi and Priestley, 2005), and the rise of the under-thrust Arabian plate after slab break-off (Bottrill et al., 2012). Most of the plateau lies within the Eurasian plate (Agard et al., 2005). There is presently little internal strain within the plateau, contrasting with continued thrusting and crustal thickening in the lower elevation parts of the Zagros, and in the Alborz mountains on the north side of the plateau (Jackson et al., 1992; Vernant et al., 2004). Initial collision may have begun at $\sim 25\text{-}35 \text{ Ma}$ (Allen and Armstrong, 2008; McQuarrie and van Hinsbergen, 2013) and continues to the present day. The GPS-derived velocity field shows roughly 20 mm/yr of north-south convergence (Vernant et al., 2004). The thickness of the lithosphere under Iran now exceeds 200 km adjacent to the Zagros suture (Priestley et al., 2012), thinning towards the margins of the collision zone.

97 Crustal thickness is likewise variable, from ~40-60 km (Seber et al., 2001; Taghizadeh-Farimand et al.,
98 2015). From these three studies the crustal thickness is around 50 km in the Dehaj study area, and the
99 lithosphere is likely to be 220 ± 50 km thick.

100

101 Today's Iranian crust is built around the Central Iranian Microcontinent (CIM), widely
102 considered to be part of a rifted terrane of Gondwanan origin called Cimmeria (Şengör et al., 1988).
103 The western parts of Iran, on the western margin of the CIM, contain the Urumieh-Dokhtar Magmatic
104 Arc (UDMA) and the Sanandaj-Sirjan Zone (SaSZ), which are crustal belts affected by Neo-Tethyan
105 subduction processes from the Jurassic to the present (Hassanzadeh and Wernicke, 2016). The SaSZ is
106 a Gondwana-derived terrane accreted to the rest of Iran during Palaeozoic Palaeo-Tethyan subduction
107 (Stampfli and Borel, 2002), and it preserves a modest record of Neo-Tethyan subduction-related
108 magmatism, particularly during the Jurassic and Eocene (Deevsalar et al., 2017). In contrast, the
109 UDMA had an upsurge in magmatic activity during the Paleocene - Eocene as a result of Neo-Tethyan
110 slab roll-back, tearing and back-arc extension (Verdel et al., 2011; Deevsalar et al. 2018). A
111 comparatively quiescent phase occurred during the Oligocene and much of the Miocene, contemporary
112 with the developing Arabia-Eurasia collision (Chiu et al., 2013; Kaislaniemi et al., 2014). In the last
113 few million years, magmatism has been abundant in eastern Anatolia, the Caucasus and NW Iran,
114 probably related to the thin lithosphere reported from this area (Keskin, 2003; Zor et al., 2003).
115 Otherwise, scattered magmatic centres are developed across the collision zone, up to 700 km from the
116 original Bitlis-Zagros suture (Shabanian et al., 2012; Kheirkhah et al., 2015; Pang et al., 2012).
117 Volcanism in the Dehaj region, in the southwest of the UDMA (Figure 1), is part of this late Cenozoic
118 magmatism. The Dehaj rocks are underlain by Neogene and Eocene volcanic and intrusive material
119 (Soheili, 1981).

120

The cause of this Late Cenozoic collision zone magmatism is debated. Major geodynamic re-organisations involving breakoff of the subducted Neo-Tethyan oceanic slab and/or detachment of the lower lithosphere have been proposed (Pearce et al., 1990; Keskin, 2003; Omrani et al., 2008). In support of this argument, detached slabs have been interpreted from tomographic images (Hafkensheid et al., 2006), corroborating predicted and observed uplift patterns (Francois et al., 2014; Magni et al., 2017). There is some debate over the specific timing of slab breakoff during the Late Cenozoic (Chiu et al., 2013). However, magmatism has seemingly persisted for several to tens of millions of years following break-off, prompting Kaislaniemi et al. (2014) to propose small scale sub-lithospheric convection as an additional trigger for magmatism. They argued that hydrated lithospheric mantle would be prone to dripping into the underlying asthenosphere, causing convection and melting of upwelling asthenosphere and potentially the down-going lithosphere. Allen et al. (2013) and Sugden et al. (2019) also proposed that melting within the mantle lithosphere could be triggered by the breakdown of hydrous phases such as amphibole.

2.2 The origin of adakitic magmas

As noted in Chapter 1, rocks of adakitic affinity are an important component of Late Cenozoic collision magmatism. These have been identified widely across Iran, with petrogenetic interpretations split between slab melting (e.g. Omrani et al. 2008), melting of crustal lithologies (e.g. Lechmann et al., 2018) and fractional crystallisation of more mafic precursor magmas (e.g. Richards et al. 2012). Adakites are defined as intermediate to felsic volcanic or intrusive samples typically containing high SiO_2 , Al_2O_3 , Na_2O , Sr, Sr/Y and La/Yb, low K_2O , MgO, $\text{K}_2\text{O}/\text{Na}_2\text{O}$, Y and Yb (e.g., Defant and Drummond, 1990; Defant et al., 1991; Martin, 1999; Castillo, 2012). There are high- (>60 wt.% SiO_2) and low-silica (<60 wt.% SiO_2) varieties, with higher MgO, Ni and Cr in low-silica adakites,

145 interpreted as a proxy for interaction between adakitic magmas and mantle wedge peridotite (Martin &
146 Moyen, 2003; Martin et al., 2005). At least some adakites or ‘adakite-like’ rocks are associated with
147 large porphyry Cu-Mo deposits (e.g., Li et al., 2011; Richards et al., 2012) and adakites may be
148 chemically similar to Archaean felsic crust (e.g., Martin, 1999; Martin et al., 2005; Hastie et al., 2015).
149 Various authors initially proposed that modern adakites are generated by the melting of young, hot
150 oceanic crust in subduction zones or at slab edges (e.g., Defant and Drummond, 1990; Peacock et al.,
151 1994).

152

153 Following the initial popularity of the slab melting model, other petrogenetic processes have
154 been proposed to produce adakitic geochemical signatures (Moyen, 2009; Castillo, 2012; Richards et
155 al., 2012). For example, the suppression of plagioclase fractionation in hydrous magmas causes Sr to
156 behave incompatibly, increasing Sr/Y ratios to higher values than in typical arc rocks as magmas
157 evolve (Richards, 2006; Richards 2011). Garnet fractionation produces depleted heavy rare earth
158 element (REE) values and hence high light/heavy (LREE/HREE) ratios, because of their partitioning
159 behaviour in the mineral ($D_{\text{LRR}} < D_{\text{HREE}}$) (Kay et al., 1987; Macpherson et al., 2006; Rodriguez et al.,
160 2007; Chiaradia et al., 2009). Amphibole fractionation also produces low Y and high Sr/Y values but
161 should be distinguishable from garnet involvement as the MREE partitioning behaviour varies between
162 the two (amphibole $D_{\text{MREE}} > D_{\text{HREE}}$, garnet $D_{\text{MREE}} < D_{\text{HREE}}$) (Richards and Kerrich, 2007; Rooney et al., 2011;
163 Davidson et al., 2013; Chen et al., 2016). Melting of under-plated or in-situ mafic lower crust
164 containing garnet and/or amphibole can produce similar overall trace element patterns to pristine slab
165 melts unaffected by mantle interaction (e.g. Atherton and Petford, 1993; Pe-Piper and Piper, 2001;
166 Chung et al., 2003; Guan et al., 2012; Xu et al., 2015).

167

168 Adakite-like rocks cannot be treated in isolation when determining their petrogenesis. For
169 example, the presence of contemporary mafic magmatism indicates concurrent mantle melting
170 processes which may rule out the possibility of pristine slab melts reaching crustal levels. At Dehaj, we
171 have identified that past work has not involved detailed analysis of a suite of mafic lavas that appear to
172 be contemporary to the known rocks of adakitic affinity. Though based on a relatively small new
173 sample set, we re-open the debate over the origin of magmatism at Dehaj, and the origin of adakites
174 more generally in Iran.

175

176 **3. Sample descriptions and analytical methods**

177

178 We collected 13 samples from the vicinity of Dehaj in Kerman Province, SE Iran (Figure 2) for
179 whole rock elemental and isotopic geochemistry. Shaker Ardakani (2016) described trachyandesitic to
180 trachydacitic subvolcanic ‘domes’ generally of felsic composition, but also containing more mafic
181 enclaves or distinct more mafic facies with evidence for mingling and/or mixing of two distinct magma
182 types, and which they included in their geochemical analyses. However, hawaiite to trachyandesite
183 lavas also occur in the area, which this paper demonstrates are geochemically similar to the more mafic
184 intrusive rocks. Eight of our samples are from these lavas which are found to the west of the Kuh-e-Aj-
185 Bala dome (~51-61 wt.% SiO₂). These lavas (Figure 3) were initially mapped as Pliocene-Quaternary
186 (Soheili, 1981), but later assumed to be of Upper Miocene – Pliocene age based on local field
187 relationships (Omrani et al., 2008). Five samples are from the subvolcanic domes or ‘stock’ of Kuh-e-
188 Aj-Bala (Figure 3). Pang et al. (2016) used secondary ionisation mass spectrometry to determine ages
189 of between 1.15 ± 0.03 and 2.15 ± 0.08 Ma (2σ) for three samples from this location. Though we
190 would prefer to have Ar/Ar geochronology for the lavas, the geochemical similarity of previously
191 analysed less evolved samples from the domes and our new lava analyses lends weight to the

192 contemporaneous nature of intrusive and extrusive magmatism (see Results, below; Table 1) (Shaker
193 Ardakani, 2016). In summary, the intrusive and extrusive rocks analysed here formed within the last
194 several million years, and they come from a restricted geographic range (Figure 2) which makes it
195 important to consider their petrogenesis in relationship to each other.

196

197 For full analytical methods for major and trace element determinations, see Neill et al. (2013),
198 which includes full trace element standard analyses run at the same time as this sample set. Sample
199 preparation was conducted at Durham University. X-ray fluorescence analysis for major elements was
200 done at the University of Leicester using a PANalytical Axios Advanced spectrometer, and trace
201 element solution chemistry at Durham University used a Thermo X2 inductively coupled plasma mass
202 spectrometer (ICP-MS). During the Dehaj sample run, Standard W2 (n = 15) gave first relative
203 standard deviations of 5% or better for all transition metals (10% for Sc, 12% for Cr, 6% for Ni), the
204 large ion lithophile elements (LILE), high field strength elements (HFSE) and the REE; excepting 7%
205 for La, 6% for Ce. Results are presented in Table 1, including sample locations. Radiogenic isotope
206 analysis at the Durham Geochemistry Centre took place as reported fully in Kheirkhah et al. (2013).
207 Nd-Sr radiogenic isotopes were analysed on a Thermo Neptune Multi-Collector ICP-MS. During the
208 Dehaj run, Sr blanks averaged 88 pg (n = 6). Standard NBS987 gave a mean $^{87}\text{Sr}/^{86}\text{Sr}$ of $0.710263 \pm$
209 12 (2σ , n = 12, minimum uncertainty 16 ppm) [corrected to a preferred value of 0.710240]. Nd blanks
210 averaged 10 pg (n = 6). The Sm-doped and un-doped versions of the J&M standard gave a mean
211 $^{143}\text{Nd}/^{144}\text{Nd}$ of 0.511099 ± 4 (2σ , n = 26, minimum uncertainty 7 ppm) [corrected to a preferred value
212 of 0.511100]. Results are presented in Table 2.

213

214 **4. Results**

215

216 4.1. Petrography

217

218 The lavas have plagioclase, rare olivine, and clinopyroxene phenocrysts, set in a fine-grained to
219 glassy groundmass (e.g. Figure 4a). Some of the olivine crystals have dark red brown iddingsite rims.
220 The groundmass contains plagioclase, olivine, clinopyroxene, rare K-feldspar and opaque minerals.
221 There are a few xenocrysts of quartz in sample DJ25.5, with reaction rims surrounding pyroxene
222 needles in the groundmass, indicating interaction of more mafic magma with a more silicic component.
223 Rare accessory phases include titanite and apatite. Some samples contain glomerocrysts, for example in
224 DJ2.1 (Figure 4b). The more acidic samples also have porphyritic textures, both in the lavas and
225 intrusive rocks (Figure 4c). The phenocrysts in the intrusive samples are mostly plagioclase (andesine,
226 with common zoning and rarer sieve texturing), with subordinate idiomorphic–hypidiomorphic green to
227 brown hornblende, rare tabular K-feldspar (sanidine) and quartz. Hornblende phenocrysts are mostly
228 green and may be strongly oxidized or have iron oxide rims. Quartz phenocrysts are rounded. The
229 groundmass in the intrusive samples is mainly composed of feldspars (sanidine and plagioclase), quartz
230 and hornblende, with apatite, biotite and iron oxides as accessory minerals.

231

232 4.2. Elemental geochemistry

233

234 New geochemical results are presented in Table 1, with associated diagrams supplemented by
235 the analyses of intrusive rocks by Pang et al. (2016) and Shaker Ardakani (2016). We begin by
236 segregating samples by intrusive vs extrusive nature (lava vs dome) and author but identify two
237 magmatic series which form a basis for later diagrams and discussion (Figure 5). Pang et al. (2016)
238 focus entirely on evolved dome samples, whereas Shaker Ardakani (2016) included more mafic facies

239 or enclaves from within the domes. The earlier published dataset of Omrani et al. (2008) is equivalent
240 to Pang et al. (2016) and hence omitted here.

241

242 Most of our samples have very low loss on ignition, but values of 1.9-2.8 % were returned for
243 mafic lavas DJ6.1, DJ25.1 and DJ25.5, those which contained the highest Sr (>3000 ppm) of our
244 sample set. At this modest loss on ignition we do not expect disturbance of immobile element
245 concentrations and ratios, but very high Sr perhaps indicates accumulation of plagioclase feldspar and
246 subsequent alteration to clay minerals, or accidental incorporation of Sr-rich amygdales, despite
247 screening during sample preparation. We highlight these samples on relevant diagrams and are mindful
248 not to involve them in discussion where appropriate.

249

250 Our extended dataset, including the lavas, demonstrates a broad compositional spectrum in the
251 sample set from hawaiite lava (~51 wt.% SiO₂) to rhyolitic intrusive rocks (~72 wt.% SiO₂), crossing
252 the alkaline to sub-alkaline divide (Figure 5a). The most mafic lava has MgO of ~6%, with 140 ppm Ni
253 and 320 ppm Cr (Table 1), not clearly representing a primary magma. When shown on a K₂O vs. SiO₂
254 plot the sample set splits into two more obvious magmatic series (Figure 5b). First, the hawaiite to
255 trachyandesite lavas and more mafic of the intrusive rocks form a high-K trend, whereas most of the
256 remaining intrusive samples fall in a medium-K trend. A few samples, including both intrusive rocks
257 and lavas, are intermediate between the two trends and are here termed ‘transitional’. All samples have
258 a sodic character, however, with NaO > K₂O (Table 1). An immobile element Th-Co diagram (Hastie
259 et al., 2007) rules out sub-solidus mobilisation of K₂O as responsible for the high- and medium-K
260 trends: one trend consists of shoshonitic basalts through to dacites (the high-K trend), and one more
261 clustered set of points consists almost exclusively of evolved calc-alkaline compositions (the medium-
262 K trend) (Fig. 5c). The same transitional samples plot between the two trends on the Th-Co plot. On

263 some Harker plots (Fig. 5d-h), such as for MgO, Sr and Y (Figures 5d, g, h), the series overlap, falling
264 on the same overall trends. For example, there are general trends between Sr and Y and SiO₂,
265 suggestive of fractional crystallisation involving plagioclase, and with Y behaving compatibly
266 throughout the evolution of the two magma series. The plots of Ni and Cr vs SiO₂ and of Rb vs SiO₂ do
267 show the distinction between the series, with the medium-K series always having much lower Ni and
268 Cr concentrations and Rb behaving incompatibly in the high-K series but with $D \approx 1$ in the medium-K
269 samples (Figures 5e, f).

270

271 The intrusive, medium-K samples have high Sr/Y such that they fall within the adakitic field on
272 a Sr/Y v Y plot, corroborating previous findings of adakitic compositions (Figure 6a). As such, from
273 now on we refer to these samples as belonging to the ‘adakite’ series. The high-K samples are
274 demonstrably not adakitic, which is a new finding for this area, including the three high-Sr lavas
275 (Figure 6a). There is a rough correlation for both series between Sr/Y ratios and SiO₂ (Figure 6b),
276 indicating that fractionation may play a role in generating or maintaining the high Sr/Y signatures, as
277 implied by the Sr v SiO₂ and Y v SiO₂ plots (Figure 5). A similar pattern is shown on a plot of La/Yb_{CN}
278 v Yb_{CN} (CN = chondrite-normalised, values from McDonough & Sun, 1995; Figure 6c), where the
279 high-K series does not plot within the adakitic field. There is a correlation in the adakite series between
280 La/Yb_{CN} and SiO₂, implying a role for fractionation in maintaining the adakitic signature. The
281 transitional samples which fall between the two series accordingly lie on the divide between adakitic
282 and non-adakitic compositions on both plots. Of the adakitic samples, SiO₂ is >60 wt.%, MgO
283 generally <3 wt.%, Ni and Cr <20 ppm and Sr >500 ppm even in the least evolved samples. As such,
284 these rocks classify as high silica adakites (Martin & Moyen, 2003).

285

286 On a chondrite-normalised diagram (Fig. 7a), the high-K hawaaitite-trachyandesite series has the
287 highest overall abundance of REEs, with moderately steep REE patterns ($\text{La/Yb}_{\text{CN}} = 12\text{-}25$) and around
288 100 to 200 times chondrite for the light REE and about 10 times chondrite for the HREE. Slight
289 negative Eu anomalies are present (geometric $\text{Eu/Eu}^* = 0.82\text{-}0.93$) implying a role for plagioclase
290 fractionation in petrogenesis. The adakite series has lower REE concentrations with steeper patterns
291 compared to the high-K series ($\text{La/Yb}_{\text{CN}} = 21\text{-}89$) and a greater spread of HREE depletion between 1-9
292 times chondrite. The Eu/Eu^* ratios vary between 0.82 and 1.18, perhaps linked to the generally
293 intrusive nature of the suite and the potential for sampling of both plagioclase-depleted and plagioclase-
294 accumulative zones within the domes. The transitional samples again lie between the two series.
295 Primitive Mantle-normalised trace element plots also demonstrate that samples share some common
296 features (Figure 7b-d). These include a positive anomaly for Sr, even ignoring the samples with the
297 highest abundances, but negative Nb, Ta, P and Ti anomalies. A modest negative anomaly for Zr
298 appears in some samples which may indicate the onset of Zr saturation. The average member of the
299 high-K series is considerably more enriched in large ion lithophile elements (LILE), e.g., Sr, Ba and
300 Th; however, the high-K series contains similar K and lower concentrations of Rb relative to the
301 adakites.

302

303 4.3. Radiogenic isotope geochemistry

304

305 The five samples analysed for Sr-Nd isotopes in this study include three adakites and two high-
306 K series samples. They fall within a limited compositional range (Table 2) predominantly in the
307 depleted quadrant with respect to Bulk Earth and within the mantle array (Figure 8). Including two
308 adakitic samples of Omrani et al. (2008) and six of Pang et al. (2016), the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ranges
309 from 0.7040 to 0.7058, and $^{143}\text{Nd}/^{144}\text{Nd}$ from 0.5128 to 0.5129. Some samples fall to the right of the

mantle array at high $^{87}\text{Sr}/^{86}\text{Sr}$ but invariant $^{143}\text{Nd}/^{144}\text{Nd}$, suggestive of sub-solidus alteration as opposed to interaction with isotopically evolved continental crust. Few Eocene rocks of the UDMA have been analysed from the local area – four samples of Omrani et al. (2008) trend towards much lower $^{143}\text{Nd}/^{144}\text{Nd}$ than the Quaternary Dehaj samples (0.5125-0.5129). More widely in the collision zone, the Dehaj rocks are isotopically quite similar to lavas considered to represent melting of peridotite or pyroxenite within the lithospheric mantle across the Lesser Caucasus (Neill et al., 2015). The Dehaj rocks, including the adakites, are derived from an isotopically quite depleted source which has not interacted with large volumes of in situ old continental lithosphere or old subducted crust.

318

319 **5. Discussion**

320

321 **5.1. Petrogenesis of the high-K series**

322

The presence of high-K mafic to intermediate lavas, plus enclaves within the domes (Shaker Ardakani, 2016) indicates a mantle source for at least some of the magmatism at Dehaj. In three recent studies (Omrani et al., 2008; Shaker Ardakani, 2016 and Pang et al., 2016), authors argued that either slab or lower crustal melting was the source of adakitic magmatism in the Dehaj area. However, there was no discussion about the origins or geological implications of contemporaneous mantle-derived magmatism. Below, the petrogenetic characteristics of the high-K series are discussed. As the majority of these rocks are trachyandesitic in composition, no detailed petrogenetic modelling was attempted.

330

Firstly, amphibole has been widely implicated in the source or fractionation history of mafic lavas across the Turkish-Iranian Plateau, as a marker for subduction-modified mantle sources of hydrous magmatism (e.g. Pearce et al., 1990), and we first seek to determine its role in the petrogenesis of the

high-K series. Further, distinguishing if magmas originated within the garnet and/or spinel peridotite facies constrains the depth of melting due to the transition between these phases at ~75 km depth in the mantle. To determine the relative roles of amphibole, garnet and spinel we look to REE systematics. It is generally accepted that Dy is more compatible in amphibole relative to garnet, and vice versa for Yb (Davidson et al., 2013) whereas spinel is compatible in neither. Davidson et al. (2013) proposed that the Dy/Dy* ratio could be a proxy for residual amphibole during melting or its fractionation during magma evolution. In the high-K series there is no relationship between Dy/Yb ratios and SiO₂ (Figure 9a). On a Dy/Dy* vs Dy/Yb plot, the high-K series follows a shallow trend not consistent with amphibole fractionation or residual amphibole but showing control from garnet, probably as a residual phase given the non-relationship between Dy/Yb and SiO₂ (Figure 9a). The sample showing the least amphibole or garnet control, mafic sample B-14 of Shaker Ardakani (2016), has La/Yb of 26 and Dy/Yb of 2.1. On the plot of Sugden et al. (2019), this sample falls in the Syunik Province and Gegham Ridge fields from southern Armenia (where the lithosphere is >100 km thick with ~45 km crustal thickness). This diagram implies that the high-K series may be derived from very small volume melting (~0.1 to 1 %) just within the garnet-spinel transition zone (Figure 9b), i.e. towards the top of the mantle lithosphere beneath Dehaj.

The Zagros suture lies ~150 km to the SW and there is no documented geophysical evidence for a subducted slab within the lithospheric mantle. Any subduction-related geochemical component is therefore most likely to have been inherited from previous subduction episodes. As noted, the hawaiite-trachyandesite high-K series is LREE- and LILE-enriched, with large negative Nb-Ta anomalies on Figure 7b. The La/Nb ratio ranges from 2.2 to 4.6, indicating a source which experienced a significant contribution from a subduction-related fluid and/or melt. On a Th/Yb vs. Ta/Yb plot (Pearce, 1983), the least evolved of the high-K series samples also fall within the subduction-modified, enriched-source

shoshonitic field (Figure 10a). There is no relationship between Yb and SiO₂, so highly incompatible behaviour of Th in the evolving magma seems to be a dominant factor in this variation rather than a specific source control. An additional consideration in understanding the mantle source comes from Rb, Sr and Ba systematics. Different trends on a Rb/Sr vs Ba/Rb plot (Furman & Graham, 1999) have been interpreted in terms of residual phlogopite (high Rb/Sr ratios) or amphibole (high Ba/Rb ratios) based on the respective Kd values in these minerals (Figure 10b). The Dehaj mafic high-K samples all lie on the amphibole trend. Even excluding samples with anomalously high Sr concentrations, there is no trend between Ba/Rb ratios and SiO₂. This finding confirms the presence of residual amphibole in what is therefore a hydrated mantle source, but also confirms the finding from REE systematics that amphibole fractionation is not important. Finally, some trace element ratios help determine if sediment-related melt or fluid release was responsible for metasomatizing the mantle source of the high-K series. There is no relationship between the ratios of Th/Nb, La/Nb and Ba/Nb and SiO₂, so these are considered reflective of source and partial melting conditions and have been applied to the diagrams of Zamboni et al. (2016). Supercritical fluids related to subducting slabs tend to have enrichments in La and Ba relative to sediment melts which have higher concentrations of Th. The Dehaj high-K series plots roughly between the two trends, implying a source influenced by a combination of both fluid and partial melt (Figures 10c and 10d).

375

376 5.2. The origin of the adakite series

377

Given past disagreement over the origin of the adakite series both at Dehaj and more widely across Iran, we now review petrogenetic evidence from the new and existing adakite geochemistry at Dehaj, considering the presence of contemporary mantle-derived magmatism.

381

382 5.2.1. Deep slab melting?

383

384 There is no geochemical evidence for interaction between the least-evolved of the high-K series
385 and deeper, upwelling magmas derived from partial melting of a subducting slab. The two distinctive
386 magma series can be clearly identified on major and trace element diagrams (Figure 5). The high-K
387 series extends to low SiO₂ values and maintains high Ni and Cr concentrations which would be diluted
388 if there were significant interaction with adakitic slab melts. The adakite series, regardless of SiO₂
389 concentrations (62-71 wt.%) has consistently low MgO (<2 wt.%), Ni (<13 ppm) and Cr (<20 ppm)
390 concentrations which are characteristic of high silica adakites which have not interacted with mantle
391 peridotite on their way to the surface (Moyen, 2009).

392

393 The rock samples classed as 'transitional' are interpreted here as the product of mixing between
394 the high-K and adakitic magmas. Is it possible that these transitional samples were slab melts which
395 interacted with mantle-derived melt at mantle depths? This is unlikely for two reasons. Firstly, there is
396 field evidence in Shaker Ardakani (2016) for mingling of the high-K and adakite series magmas at sub-
397 volcanic depths. Secondly, transitional samples are confined to a narrow range of 60-62 wt.% SiO₂ and
398 do not display any evidence of fractional crystallisation in their own right. Should these transitional
399 magmas have been generated at mantle depths, shallower crustal magmatic evolution would occur in
400 that series, generating a greater spread of SiO₂ for transitional compositions. As such, we cannot
401 identify evidence from any of the erupted or emplaced rocks of interaction between slab melts and the
402 mantle or mantle-derived magmas.

403

404 Finally, at a distance of ~150 km from the Zagros Suture, and with tomographic evidence for a
405 detached Zagros slab within the asthenosphere beneath the collision zone (van der Meer et al., 2018), it

seems unlikely that any slab material is physically present within the mantle lithosphere beneath Kerman Province. We therefore rule out the hypothesis of Omrani et al. (2008).

5.2.2. Fractional crystallisation of a mafic parent?

The high-K series is the only mafic to intermediate magmatism for which there is surface evidence of contemporaneous formation with the adakites. We have no reason to suspect the presence of any other potential mafic precursors to the adakites. The high-K and adakite series are clearly geochemically distinct and the former cannot be parental to the latter. Firstly, the high-K series has a greater concentration of most incompatible elements relative to the adakites (Table 1). The high-K series also has higher $^{87}\text{Sr}/^{86}\text{Sr}$ and lower $^{143}\text{Nd}/^{144}\text{Nd}$ compared to that of many of the adakites (Figure 8). Therefore, any assimilation-fractional crystallisation model would require that the high-K series be contaminated by isotopically depleted, medium- to low-K crust. Geologically this is not a feasible scenario given the known compositions of the continental crust and of Eocene and younger magmatism in the region. Finally, there are no adakites lower than ~62 wt.% SiO_2 , and the low MgO and transition metal concentrations within the adakite series do not belong to the same geochemical trend as the most-evolved of the high-K series (Figure 5). Therefore, a fractional crystallisation origin for the adakites from this mafic, mantle-derived parental magma is not likely.

5.2.3. Melting mafic lower crust: in situ or delaminated?

Next, we consider if the adakites are derived from melting of a crustal lithology perhaps at the base of the UDMA crust. The adakites are isotopically depleted and there is no positive relationship between $^{87}\text{Sr}/^{86}\text{Sr}$ and the degree of magma evolution. Therefore, if the adakites are ultimately sourced from the

430 melting of crustal lithologies, that crust must not be ancient or otherwise isotopically evolved but
431 should have formed in the recent past. Geologically this is feasible: subduction of the southern branch
432 of the Neo-Tethys ocean led to subduction-related magmatic episodes from the Mesozoic to the early
433 Eocene, followed by the Late Eocene emplacement of within-plate magmas during Neo-Tethyan slab
434 roll-back (Verdel et al. 2011; Deevsalar et al. 2018). It is reasonable to assume that young mafic lower
435 crust, perhaps underplated, lies beneath Dehaj.

436

437 Clemens et al. (2006) demonstrated that plagioclase is stable within the crust to about 1.5 GPa, or
438 approximately 50 km. At shallower depths, the presence of residual plagioclase during partial melting
439 is likely to result in the production of low-Sr (<400 ppm) magmas (Smithies et al., 2009). In our sample
440 set and the analyses of Shaker Ardakani (2016) and Pang et al. (2016), Sr concentrations in the adakites
441 are consistently > 400 ppm, and those adakites which are least evolved, containing ~60-64 wt.% SiO₂,
442 have 800-1800 ppm Sr. This finding is a pointer towards either the complete dissolution of plagioclase
443 during partial melting at amphibolite facies, or its absence during melting within the deeper eclogite
444 facies at >50 km. It would represent an unusually high degree of melting for plagioclase to be
445 completely consumed from a crustal amphibolite which might contain many tens of % plagioclase.
446 Furthermore, as there is only evidence for small volumes of mafic, mantle-derived magma (the high-K
447 series), heat advection is unlikely to be a strong driver of extensive lower crustal melting. It is also
448 unlikely that the source would be a plagioclase-poor lithology such as a mafic to ultramafic cumulate,
449 especially as the melting temperature of such a rock would be high and therefore other plagioclase-
450 bearing lithologies would also melt out, generating low-Sr non-adakitic magmas. It seems much more
451 likely that plagioclase was absent in the source rock and that the source was therefore transitioning to,
452 or within, the eclogite facies.

453

Recent determinations indicate the UDMA crust beneath Dehaj to be approximately 50 km thick (see Section 2), on the cusp of formation of eclogite. Allowance must be made for uncertainty in crustal thickness models, so the simplest explanation for generation of adakitic magmas in this setting is that the lowermost crust partially melted during the amphibole-eclogite transition, aided by the fact this was taking place at the thermal boundary with the underlying mantle. Should the crust in reality be somewhat *thinner* than 50 km, others such as Lechmann et al. (2018) have proposed that, during the ongoing Arabia-Eurasia collision, slivers of garnet amphibole were torn from the base of the Iranian crust and tectonically interwoven with the lithospheric mantle. These crustal fragments would therefore be subjected to higher pressures and temperatures than at the Moho, converting to eclogite facies, at which point dewatering and partial melting could take place. Lechmann et al. (2018) proposed this model to explain why adakites in NW Iran, which require an eclogitic source, were generated in a region of thinner orogenic crust (~45 km) without slab melting. Lechmann et al. (2018), like us, considered a fractional crystallisation model difficult to reconcile with local geological evidence. Alternatively, it is possible that crustal thickening during collision can ‘eclogitize’ mafic lower crust, to the point where it becomes seismically less distinguishable from the surrounding mantle (e.g. Dewey et al. 1993). This model allows for the crust to be *thicker* than geophysical estimates allow, giving us more confidence that the lower crust can be a source of the adakitic series. Therefore, we have confidence that melting of eclogitic rocks took place at >50 km and at shallower depths than the beginning of the onset of the spinel-garnet transition in mantle lithologies.

473

474 5.3. Fractional crystallisation vs the strength of the adakitic signature

475

476 Regardless of the source model, we have already noted that the evolution of the adakite series
477 from ~60-72 wt.% SiO₂ includes sharp upturns in La/Yb and a steady increase in Sy/Y ratios (Figure

478 6b, d). Therefore, two iconic signatures of adakitic magmatism show a direct relationship with the
479 degree of magma evolution. On Figure 9a, there are very variable Dy/Yb ratios consistent with the
480 involvement of garnet, and that ratio becomes greater the more evolved the adakites are. This finding
481 strongly implies a garnet control during fractional crystallisation which modifies the adakite signature
482 at Dehaj. Garnet, the only major phase capable of strongly fractionating the heavy REE, must
483 fractionate at depth as no garnet is physically seen in the petrography. Secondly, although Sr behaves
484 compatibly, indicating plagioclase fractionation, the Sr/Y ratio is largely controlled by garnet
485 compatibility in Y. The narrow range of Dy/Dy* ratios in the adakites (Figure 9a) also implies no
486 strong role for amphibole during magma evolution. In short, garnet is a major player in the evolution of
487 the Dehaj adakite signature.

488

489 We also consider the geochemistry of other late Cenozoic igneous suites from across Iran and
490 adjacent areas. Several of these have previously been reported as having adakitic compositions
491 (Jahangiri, 2007; Omrani et al., 2008; Khodami et al., 2009; Shafiei et al., 2009; Richards et al., 2012;
492 Shabanian et al., 2012; Azizi et al., 2014; Asadi et al., 2014; Ghorbani et al., 2014; Zarasvandi et al.,
493 2015; Lechmann et al., 2018), mainly from middle and Late Miocene centres. The youngest
494 (Quaternary) centres, and some Late Miocene centres, lack detailed reports of adakitic compositions
495 (e.g. Kheirkhah et al., 2009; Allen et al., 2013), although evolved rocks were not always the focus in
496 some of these works, and Pang et al. (2016) and Lechmann et al. (2018) both noted several <5 Ma
497 volcanic centres in northwest Iran, Armenia and Georgia as having adakitic compositions. Late
498 Cenozoic adakitic rocks seem to be rarer in the western parts of the Arabia-Eurasia collision zone,
499 where the crust is thinner. As noted, some have interpreted the adakitic signature as indicating melting
500 of the Tethyan oceanic slab (e.g., Omrani et al., 2008; Khodami et al., 2009; Azizi et al., 2014), versus

501 fractionation of mantle-derived melts (e.g., Richards et al., 2012), or melting of in-situ or delaminated
502 lower crust (Shafiei et al., 2009; Pang et al., 2016; Lechmann et al., 2018).

503

504 In looking more closely at the geochemistry of such suites from Dehaj and across Iran, a
505 common theme appears to be the systematic variation in intra-REE ratios such that each of the suites
506 has a geochemical signature dominated by *either* amphibole or garnet. As already noted, the Dehaj
507 samples appear to have experienced garnet fractionation during magmatic evolution, and this is picked
508 out by the steep trend on a Gd/Dy vs Dy/Yb plot (Figure 11a). However, other suites with amphibole-
509 dominated trends are picked out in narrower ranges of Gd/Dy forming roughly linear arrays relative to
510 Dy/Yb such as the Kerman dataset of Shafiei et al. (2009), given that amphibole has $D_{Dy} < D_{Gd}$
511 (Davidson et al., 2013). The garnet association is apparent not just at Dehaj, but also Tabriz (Jahangiri,
512 2007) and Meshkan (Shabanian et al., 2012). A trend more consistent with amphibole involvement is
513 present in the data from different centres in Kerman Province, SE Iran (Shafiei et al., 2009; Zarasvandi
514 et al., 2015). Some datasets are not conclusive, perhaps because several centres are grouped together, as
515 is currently the case for centres near Isfahan (Khodami et al., 2009), or the compositional range is too
516 small to show fractionation trends (Pang et al., 2016). Similar relationships can be determined from the
517 plot of La/Sm v Sm/Yb (Figure 11b). Whereas the Dehaj samples of our study and Pang et al (2016)
518 form an array heading towards high Sm/Yb values (i.e. a garnet signature), other suites from Iran plot
519 in more restricted fields towards the low-Sm/Yb side of the diagram, indicating amphibole control, or
520 possibly a joint amphibole/garnet effect.

521

522 5.4. Relationship to mineralization

523

524 Of all the adakitic suites described across Iran in recent years, it is notable that whilst a few are
525 clearly linked to economic porphyry copper mineralization (Shafiei et al., 2009; Zarasvandi et al.,
526 2015), most are not (Jahangiri, 2007; Omrani et al., 2008; Khodami et al., 2009; Shabanian et al., 2012;
527 Azizi et al., 2014; Pang et al., 2016; this study). We can now reveal that suites associated with
528 economic mineralization are more *generally* those where amphibole is strongly implicated in the
529 petrogenetic history (Figure 11a, b), specifically some of the centres from the Kerman region, SE Iran
530 (Shafiei et al., 2009; Zarasvandi et al., 2015). Garnet is more clearly involved in the development of
531 centres that do not contain economic deposits (including the Dehaj centre in this study), although
532 Shafiei et al (2009) noted garnet signatures (high Sm/Yb) in just two individual samples from centres
533 with economic deposits. We regard the association of amphibole signatures and mineralization, but not
534 garnet, as robust enough to merit further investigation. The La/Sm v Sm/Yb diagram has previously
535 been used for the study of magmatic rocks associated with copper and gold mineralization in the
536 Central Andes (Figure 11b; Kay and Mpodozis, 2001). Notably few Iranian samples plot within the
537 field associated with copper and/or gold mineralization in the Andes, including suites from economic
538 centres in the Kerman region (Shafiei et al 2009; Zarasvandi et al., 2015).

539

540 Why some centres should show amphibole control on MREE and HREE fractionation, and
541 others garnet control, is clearly a subject for further study. Overall crustal thickness is likely to be an
542 important factor in the depth at which magmas crystallise (Kay and Mpodozis, 2001), but unlikely to be
543 a complete explanation, given that economic and non-economic centres exist within the same region of
544 SE Iran, developed at roughly the same time. Although many of the major copper deposits in SE Iran
545 are Middle to Late Miocene age (16-5 Ma), others are Early Miocene, coinciding with barren centres
546 (see age compilations in Shafiei et al., 2009 and Zarasvandi et al., 2015). It may be that there are more
547 local controls on the depths at which magma fractionates, for example the stress conditions at the time

548 of magma transfer from the mantle through the crust (Hutton, 1982) or the extent to which metals are
549 concentrated in the magmas or scavenged from the local crust.

550

551 **6. Conclusions**

552

553 Contemporary igneous rocks from Dehaj, SE Iran, belong to two distinct magmatic series. One
554 series is entirely intrusive and is of high silica adakitic affinity and the second, comprising both
555 hawaiite to trachyandesite lavas and intrusive rocks, is of high-K, non-adakitic affinity. Some rocks are
556 geochemically transitional between the series and are proposed as hybrids formed by magma mixing
557 between the adakitic and high-K series.

558

559 The high-K rocks are derived from melting of subduction-modified lithospheric mantle in the
560 shallowest part of the garnet-spinel transition zone. The chemical signatures and degree of LREE and
561 LILE enrichment are consistent with very small degree partial melting (0.1-1 %), mirroring models for
562 other parts of the Turkish-Iranian Plateau with thick lithosphere (Allen et al., 2013; Sugden et al.,
563 2019). It is determined that the adakitic samples are unrelated to the high-K series and did not
564 experience interaction with mantle wedge peridotite. They are therefore unlikely to be partial melts
565 from a subducted slab at depths greater than that responsible for generation of the high-K series. The
566 adakites have high Sr concentrations, demanding that plagioclase was absent or breaking down during
567 partial melting. Such eclogite-facies conditions might lie at the very base of the crust or as slivers of
568 tectonically delaminated lower crust incorporated in the mantle.

569

570 Within the adakite series, increasing SiO₂ is correlated with markers of adakitic chemistry such
571 as La/Yb and Sr/Y. Garnet fractionation during magma evolution is the key agent for decreasing HREE

ratios, based on intra-REE variations. Garnet fractionation is consistent with the stalling of upwelling
magmas in deep seated magma chambers within the thickened crust in this part of the Turkish-Iranian
Plateau. We interpret several other late Cenozoic igneous suites from Iran to have involved garnet
and/or amphibole during the production of adakite-like signatures, either as a residual or fractionating
phase, and suggest that sites of porphyry mineralisation associated with these rocks tend to favour those
which involved amphibole and not garnet. As a final note, we caution for care in interpreting the source
and evolution of adakitic rocks. As has been observed here, past work has often been biased towards
felsic lithologies meaning that contemporary mafic lithologies, which play a critical role in limiting
possible adakite origins, may be overlooked.

581

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931

932 **Figure captions**

933 *Figure 1. Location map for Late Cenozoic magmatism on the Turkish-Iranian Plateau. After*
934 *Kaislaniemi et al. (2014).*

935

936 *Figure 2. (a) Geology of the Dehaj region, adapted from the 1:250,000 Anar map (Soheili, 1981) and*
937 *overlain with the main sampling locations for mafic lavas and dome samples. (b) satellite imagery from*
938 *<https://earth.google.com> (accessed 05/08/2019), showing the relationship between the older Mesozoic*
939 *to Cenozoic magmatic rocks and the two main Quaternary adakitic domes and mafic lavas.*

940

941 *Figure 3. Field photos of volcanic centres in the Dehaj region. a) DJ1-1, one of the mafic, high-K*
942 *series lava flows; b) heavily jointed intermediate lava flow DJ2-1 of the transitional samples; c) DJ25-*
943 *3 in the foreground, another high-K series mafic lava flow, with the mountain Kuh-e-Aj-Bala in the*

944 background, a large stock or dome of the adakite series. See text for full details of the different
945 magmatic series.

946

947 Figure 4. Petrography of the newly analysed samples, all with 4 mm horizontal field of view in cross-
948 polarised light. a) DJ1-1 (high-K series) showing groundmass with clinopyroxene phenocrysts,
949 plagioclase and oxides with rare iddingsite replacing olivine. b) DJ2-1 (transitional series) showing a
950 1 mm glomerocryst of clinopyroxene set in a very fine groundmass. c) DJ5-1 (adakite series) showing a
951 seriate texture dominated by plagioclase, often zoned but not sieve-textured, e.g., the crystal in the
952 centre of the image.

953

954 Figure 5. Geochemical discrimination and Harker plots for the Dehaj samples, including previous
955 analyses by Shaker Ardakani (2016) and Pang et al. (2016). a) Total alkalis vs SiO_2 . b) K_2O vs SiO_2
956 after Peccerillo & Taylor (1976). c) Th-Co for immobile elements after Hastie et al. (2007)
957 demonstrating the two main series proposed on (b) are valid. d-g) Harker plots demonstrating
958 compatible behaviour of MgO, Ni and Cr, Sr and Y and incompatible behaviour of Rb. Three high-K
959 lavas with Sr concentrations >3000 ppm have been omitted from (f).

960

961 Figure 6. Adakite vs typical arc rock discrimination diagrams from Moyen (2009) alongside ratio-
962 silica Harker plots for adakite-discriminating trace elements, with samples from this work plus Shaker
963 Ardakani (2016) and Pang et al. (2016). a) Sr/Y vs Y. b) Sr/Y vs SiO_2 . Having identified the
964 adakitite/medium-K and non-adakitite/high-K series on earlier diagrams, the next two diagrams merge
965 data sources and consistently use this new distinction. c) La/Yb_{CN} vs Yb_{CN} (CN = chondrite-
966 normalised). d) La/Yb_{CN} vs SiO_2 . The cut-offs for Sr/Y and La/Yb are from Defant and Drummond
967 (1990), based on Sr/Y and La/Yb (not CN) = 40 and 20, respectively.

968

969 *Figure 7. Multi-element plots for Dehaj incorporating this work plus Shaker Ardakani (2016) and*
970 *Pang et al. (2016). a) Chondrite-normalised REE plot showing all three series. b-d) Primitive mantle-*
971 *normalised spider diagrams, showing b) the high-K series compared with the average adakite; c) the*
972 *adakite/medium-K series with the average high-K series sample; d) the transitional samples showing*
973 *the other series averages for comparison.*

974

975 *Figure 8. Isotopic composition of the two series, including samples analysed for this study, Pang et al.*
976 *(2016) and Omrani et al. (2008). No transitional samples were analysed. As in previous diagrams, the*
977 *adakite series is represented by black squares, and the high-K series by open circles. See the named*
978 *papers on the diagram for comparative data sources.*

979

980 *Figure 9. a) Dy/Dy* diagram after Davidson et al. (2013) showing the enriched nature of the*
981 *source/low volume melting required to form the high-K series (circles), versus the very strong apparent*
982 *garnet control on the REE systematics of the adakite series (squares). The inset shows no relationship*
983 *between Dy/Yb and magmatic evolution for the high-K series, but there is a stronger relationship for*
984 *the adakites. b) Comparison of the high-K series with melt models for amphibole-bearing garnet and*
985 *spinel peridotite for Armenia by Sugden et al. (2019). Syunik in S Armenia represents melting within*
986 *thick lithosphere during amphibole peridotite breakdown, within the garnet-spinel transition zone;*
987 *melting in Shirak in N Armenia is typical of the more voluminous melting at shallow depths across*
988 *much of the Caucasus and E Anatolia.*

989

990 *Figure 10. a) Th/Yb vs Ta/Yb from Pearce (1983) showing the subduction-related geochemical*
991 *signature of the high-K series, with the inset demonstrating little relationship between Th/Yb and*

992 magmatic evolution. b) Rb/Sr vs Ba/Rb from Furman & Graham (1999) showing a strong residual
993 amphibole control on the high-K series. c) Th/Nb vs La/Nb and d) Th/Nb vs Ba/Nb plots from Zamboni
994 et al. (2016) are used to explore the nature of the past subduction-related materials which contributed
995 to the mantle source of the high-K series, implying a combination of slab-related melts and fluids were
996 responsible.

997

998 Figure 11. a) Gy/Yb vs Dy/Yb diagram designed to separate the steep trends related to intra-HREE
999 fractionation caused by amphibole at source or amphibole fractionation, versus garnet. b) La/Sm vs.
1000 Sm/Yb to more clearly highlight the role of amphibole at source or during fractionation. Data sources
1001 are given on the figure.

1002

1003 **Table captions**

1004

1005 Table 1. Major and trace element data for the studied rocks from Dehaj, SE Iran. HK = high K series.
1006 Transit' = transitional, i.e. samples which fall between the high K and adakite series on various
1007 Harker plots and are interpreted to have formed by mixing between the two series. Adakite = meets the
1008 definition of high silica adakite (Martin and Moyen, 2003). b.d. = below detection.

1009

1010 Table 2. Sr and Nd isotope data for the studied rocks from Dehaj, SE Iran. Values are normalised to
1011 NBS987 $^{87}\text{Sr}/^{86}\text{Sr} = 0.710240$ and J&M $^{143}\text{Nd}/^{144}\text{Nd} = 0.511110$ and are not age corrected.

1012

Figure 1

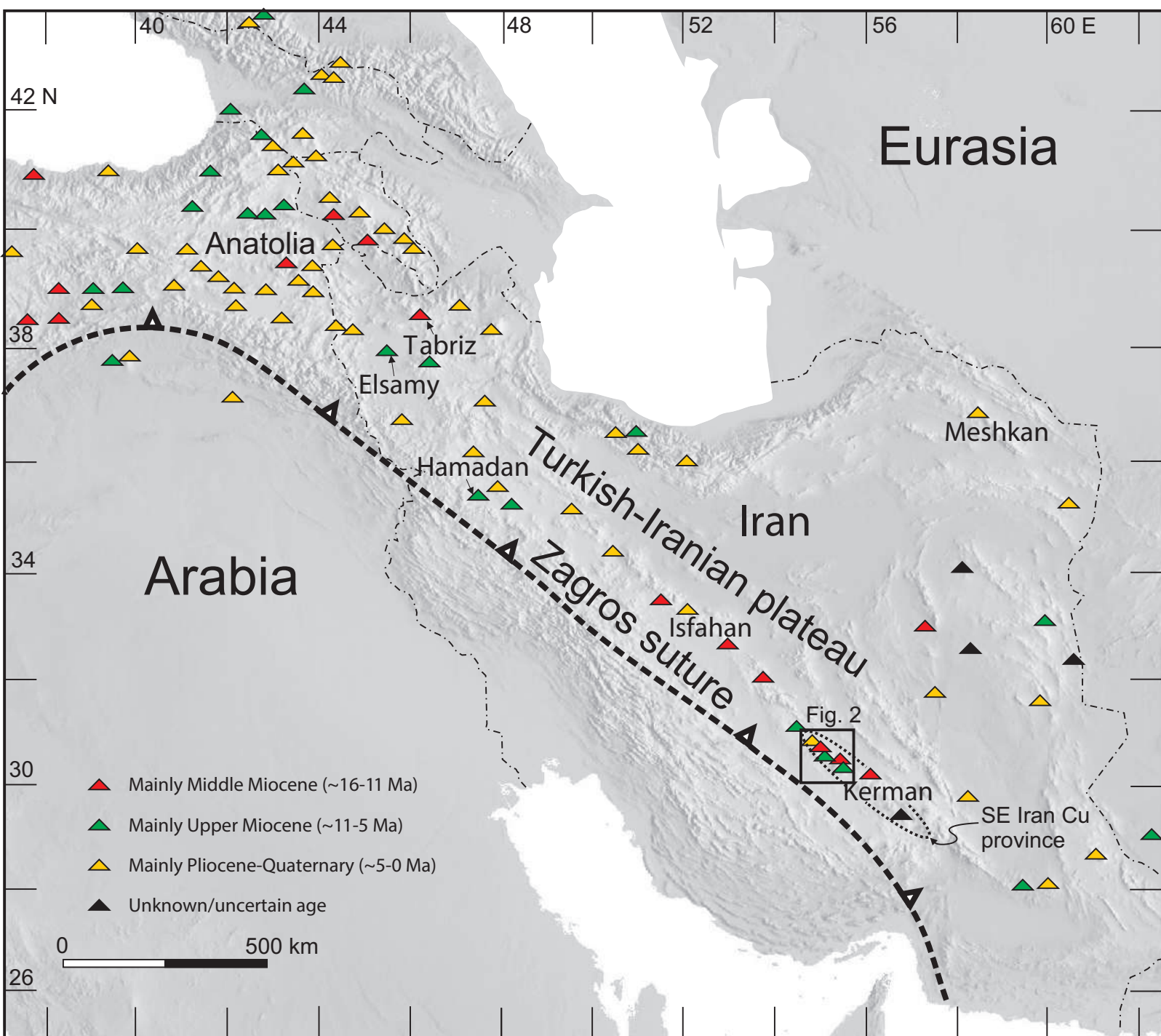


Figure 2

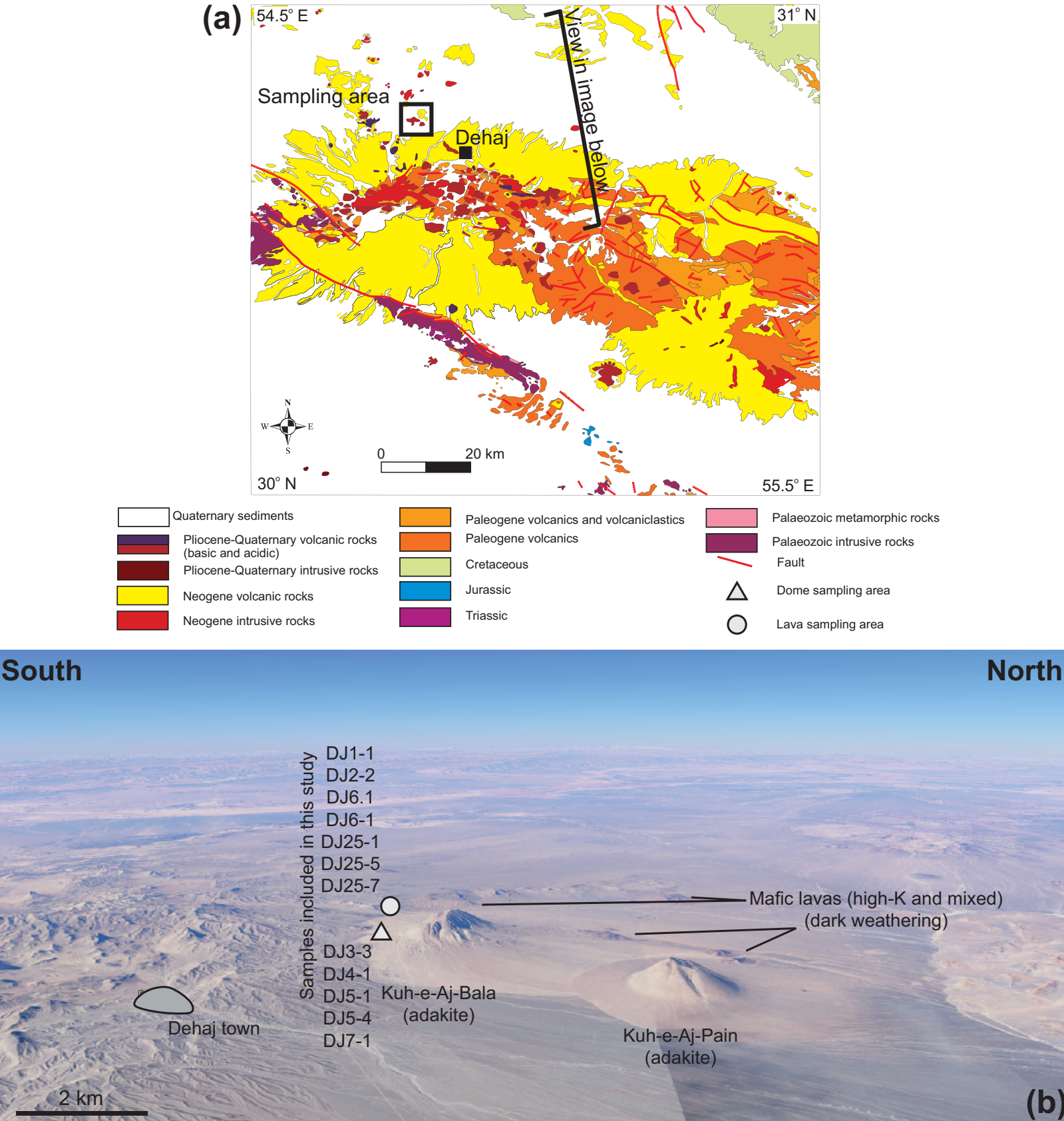


Figure 3

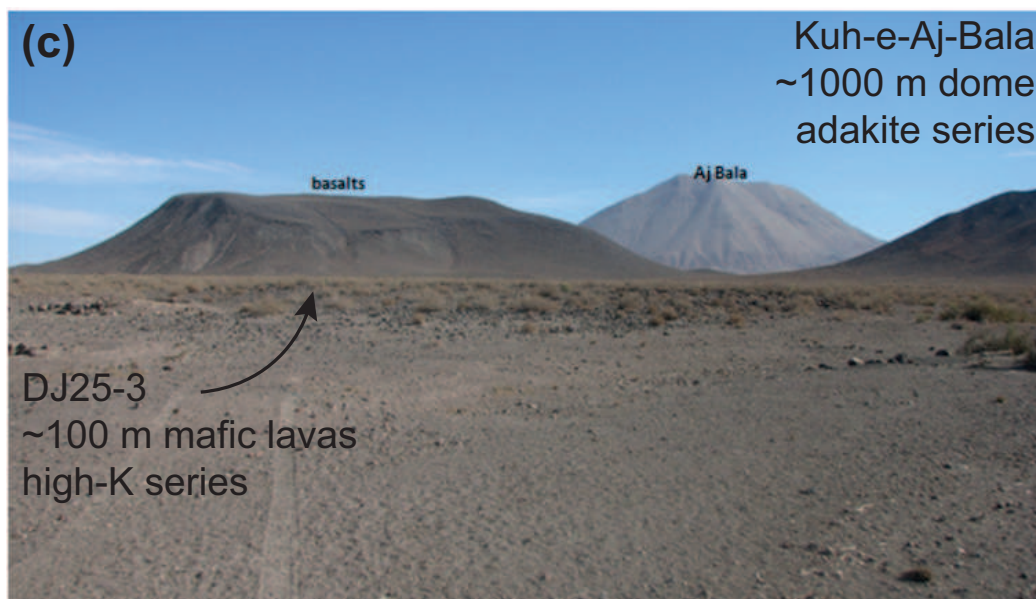
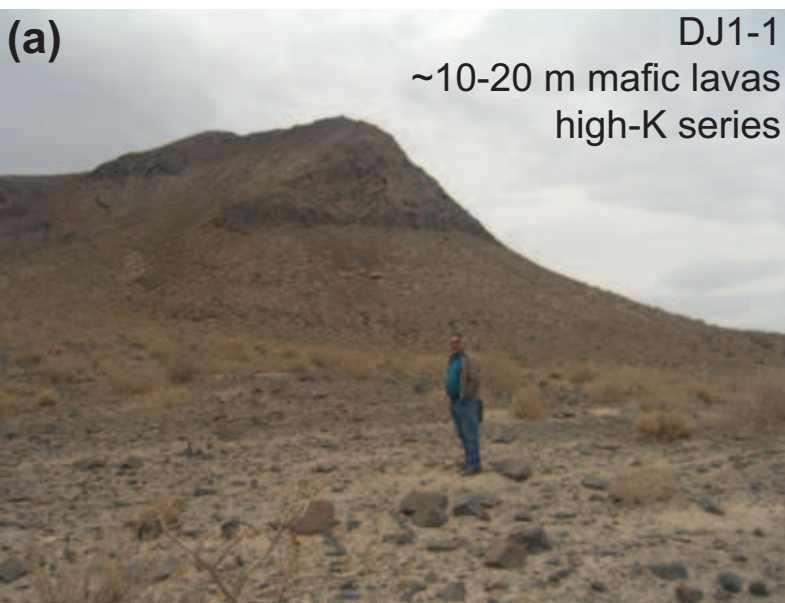
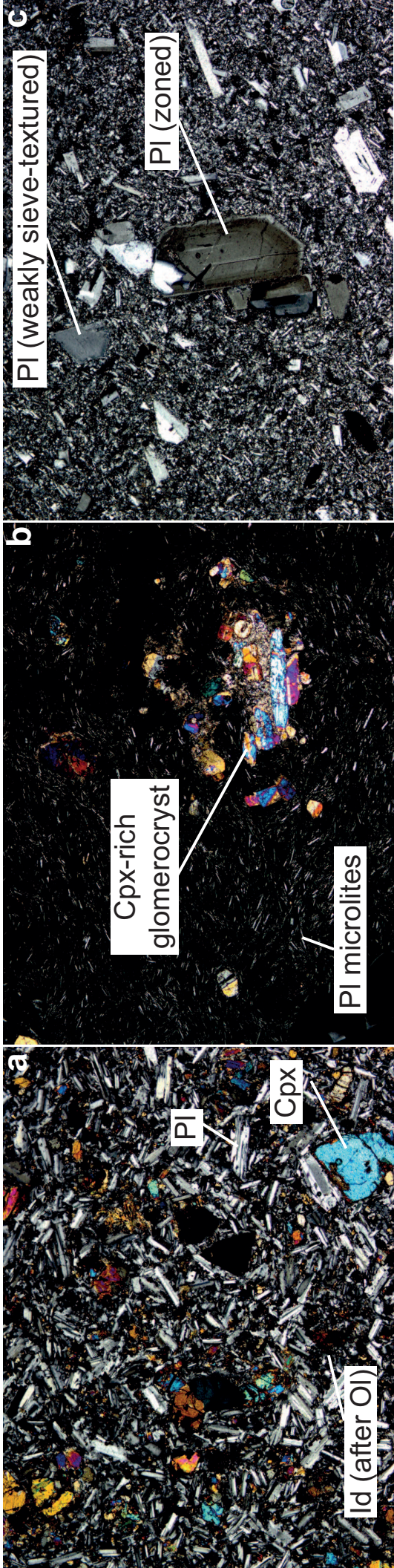


Figure 4



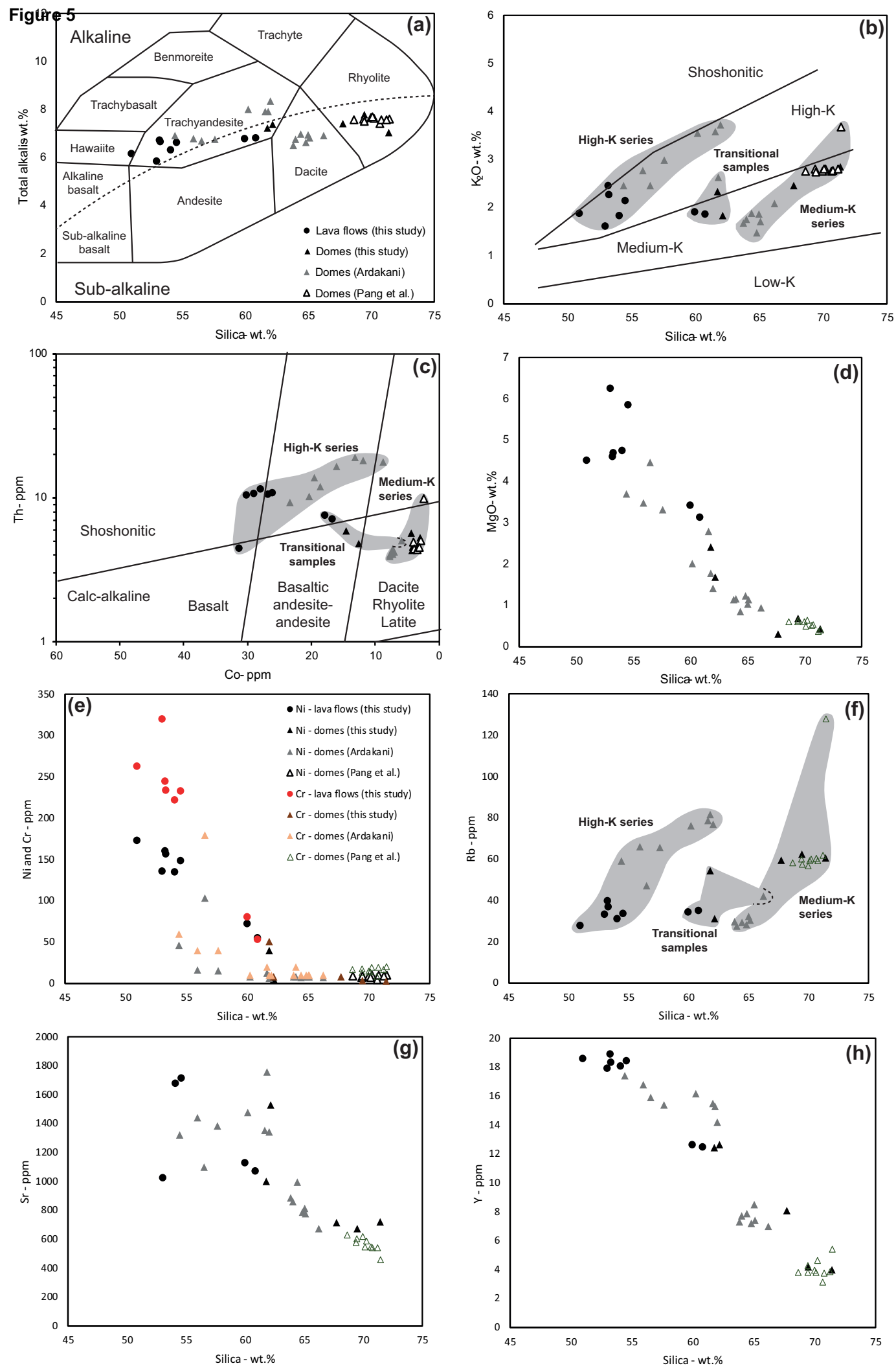


Figure 6

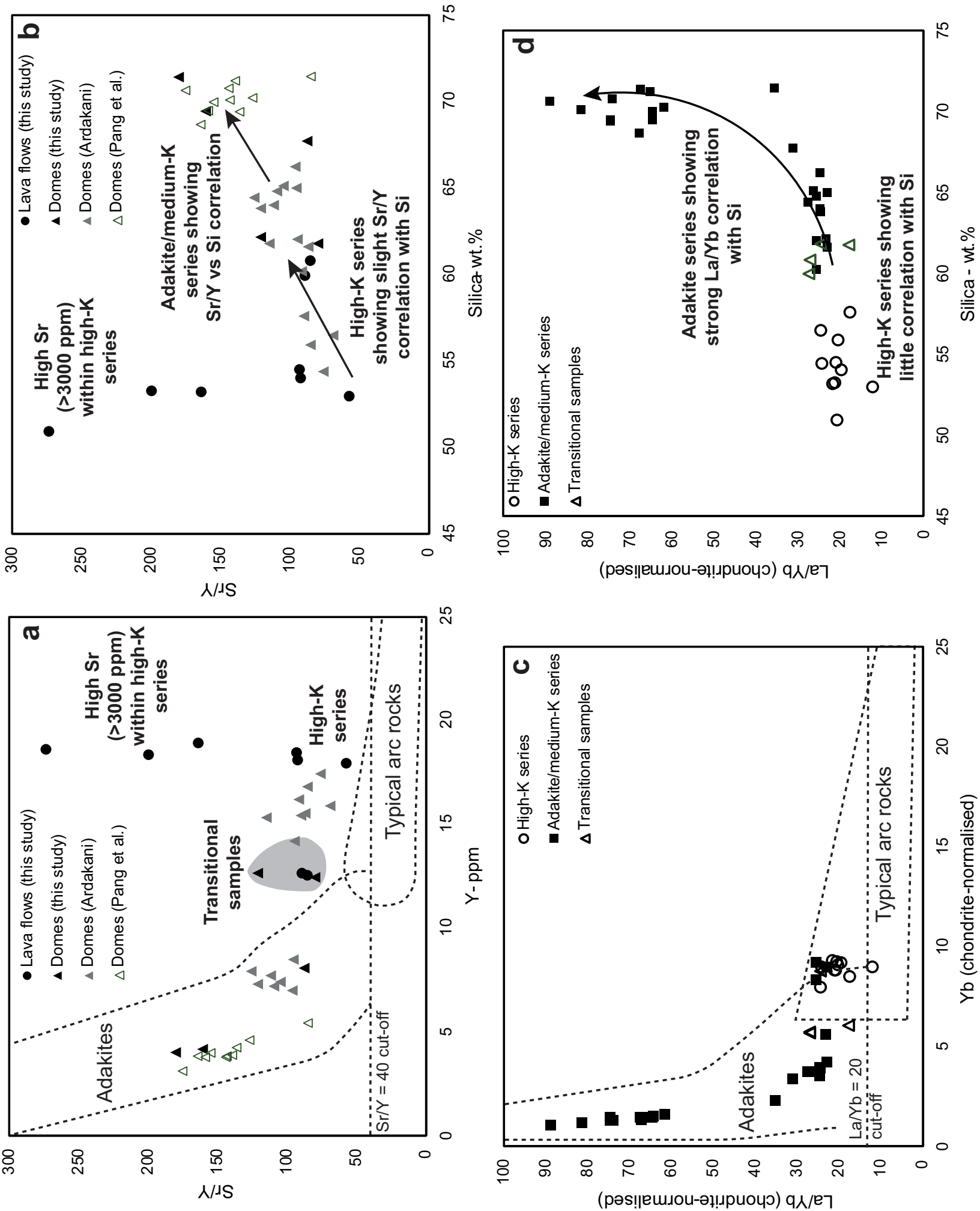


Figure 7

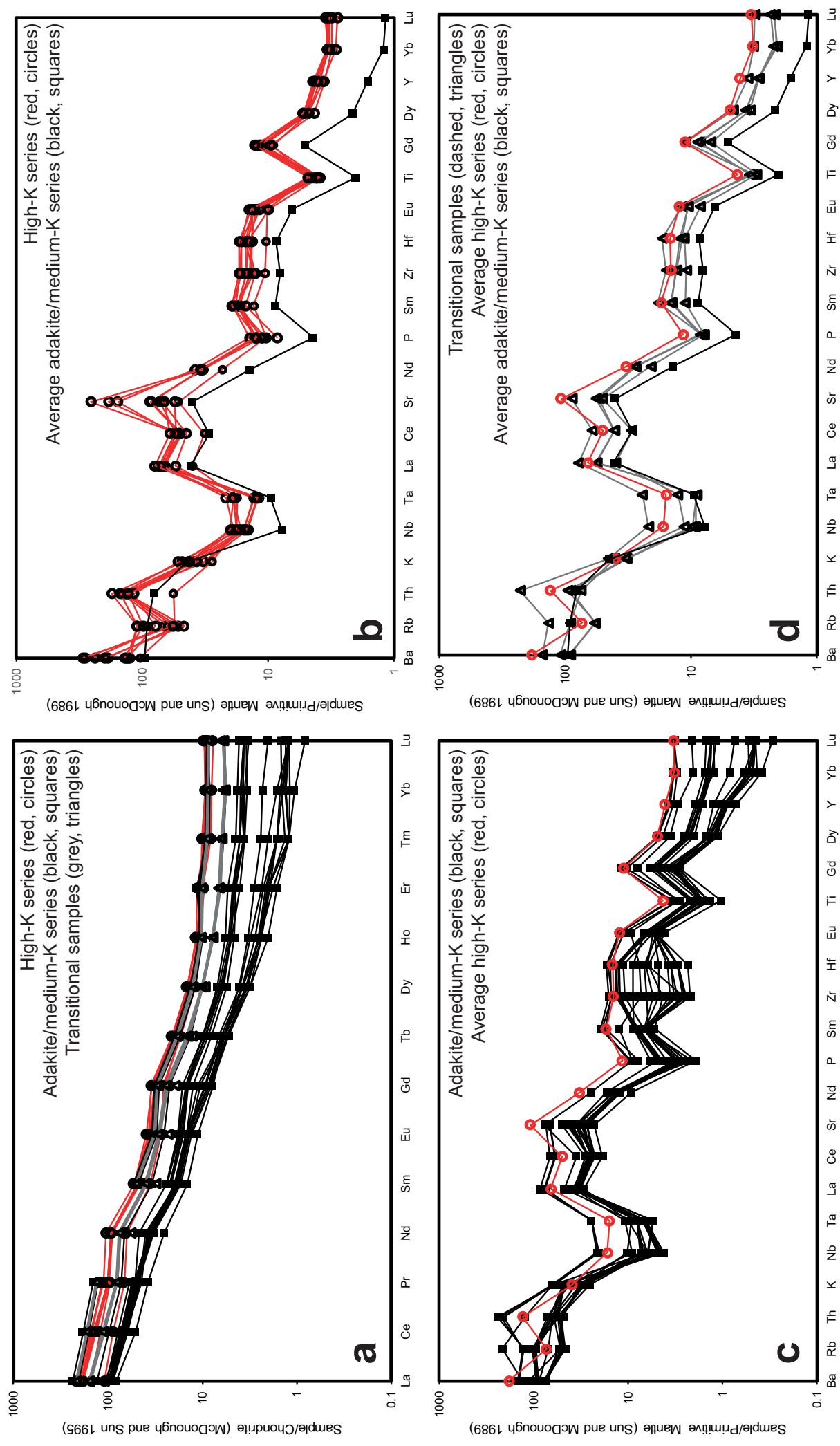


Figure 8

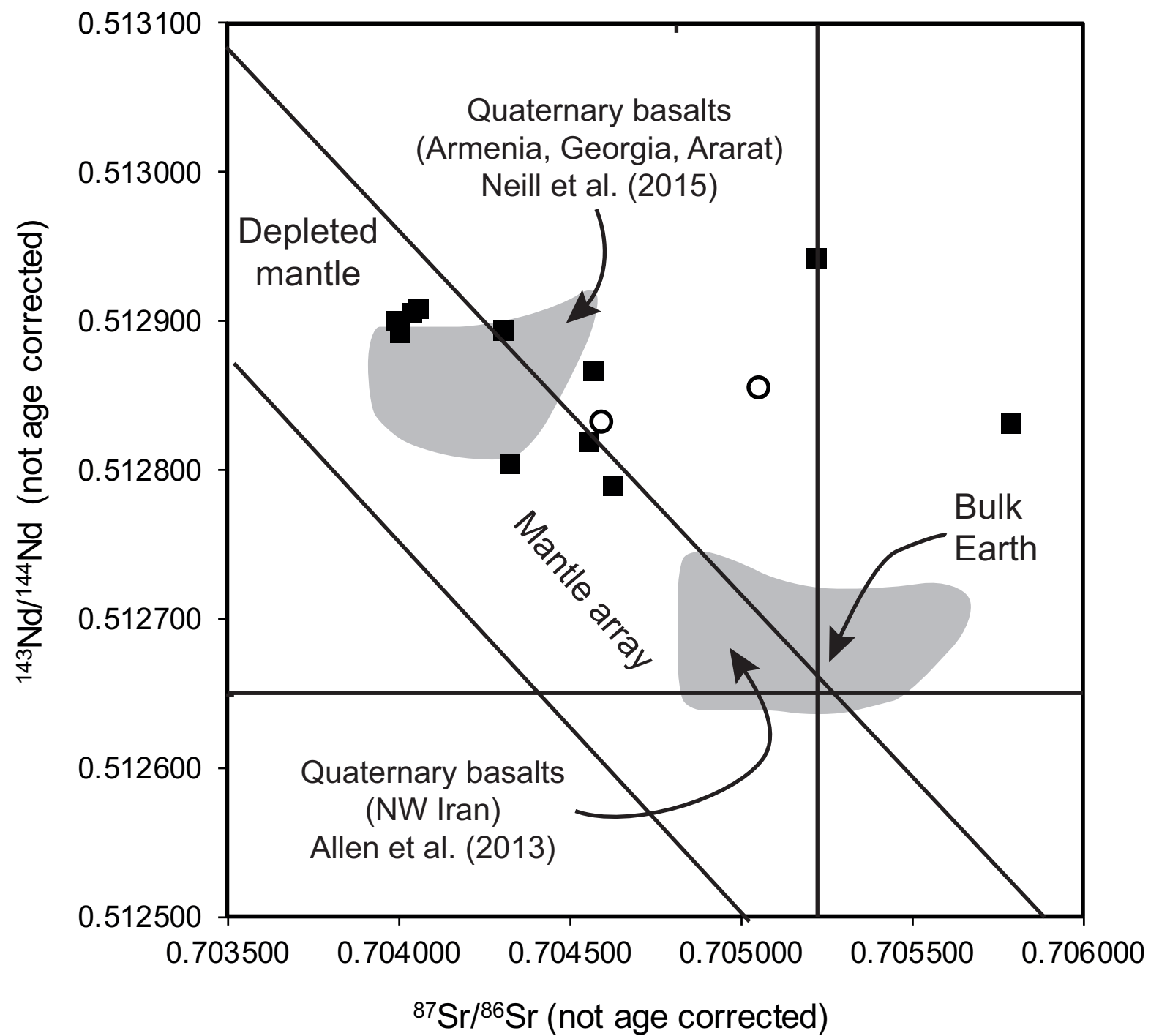


Figure 9

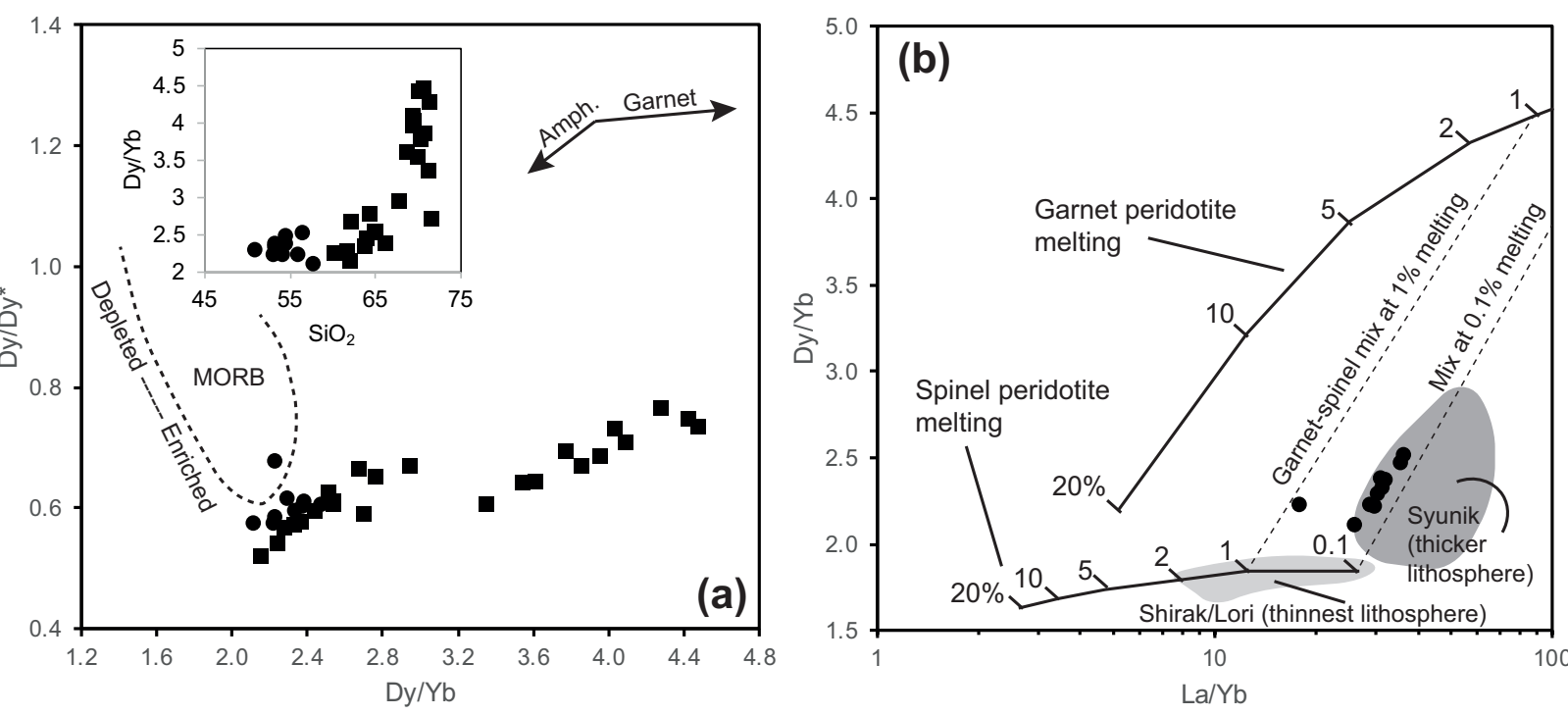


Figure 10

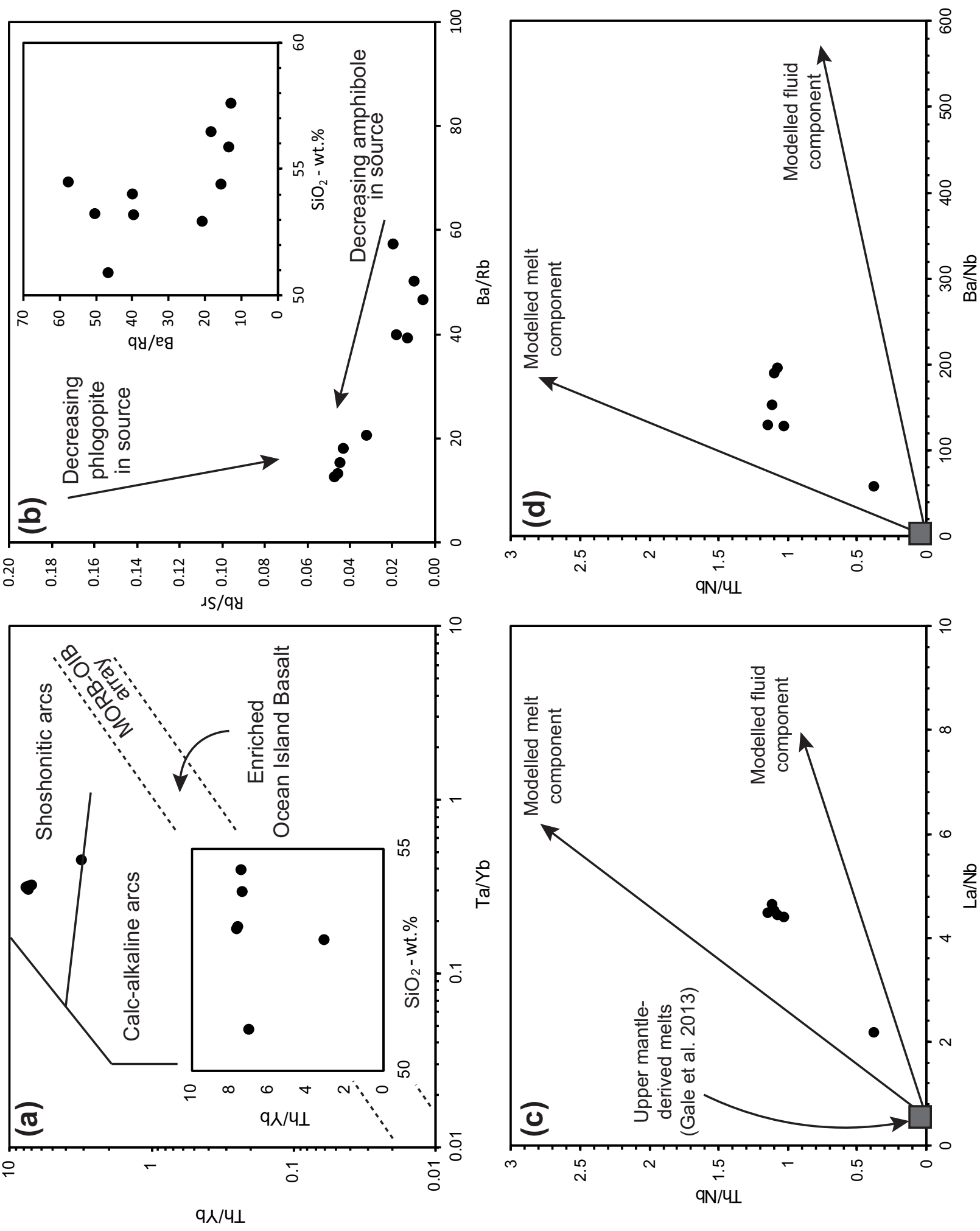


Figure 11

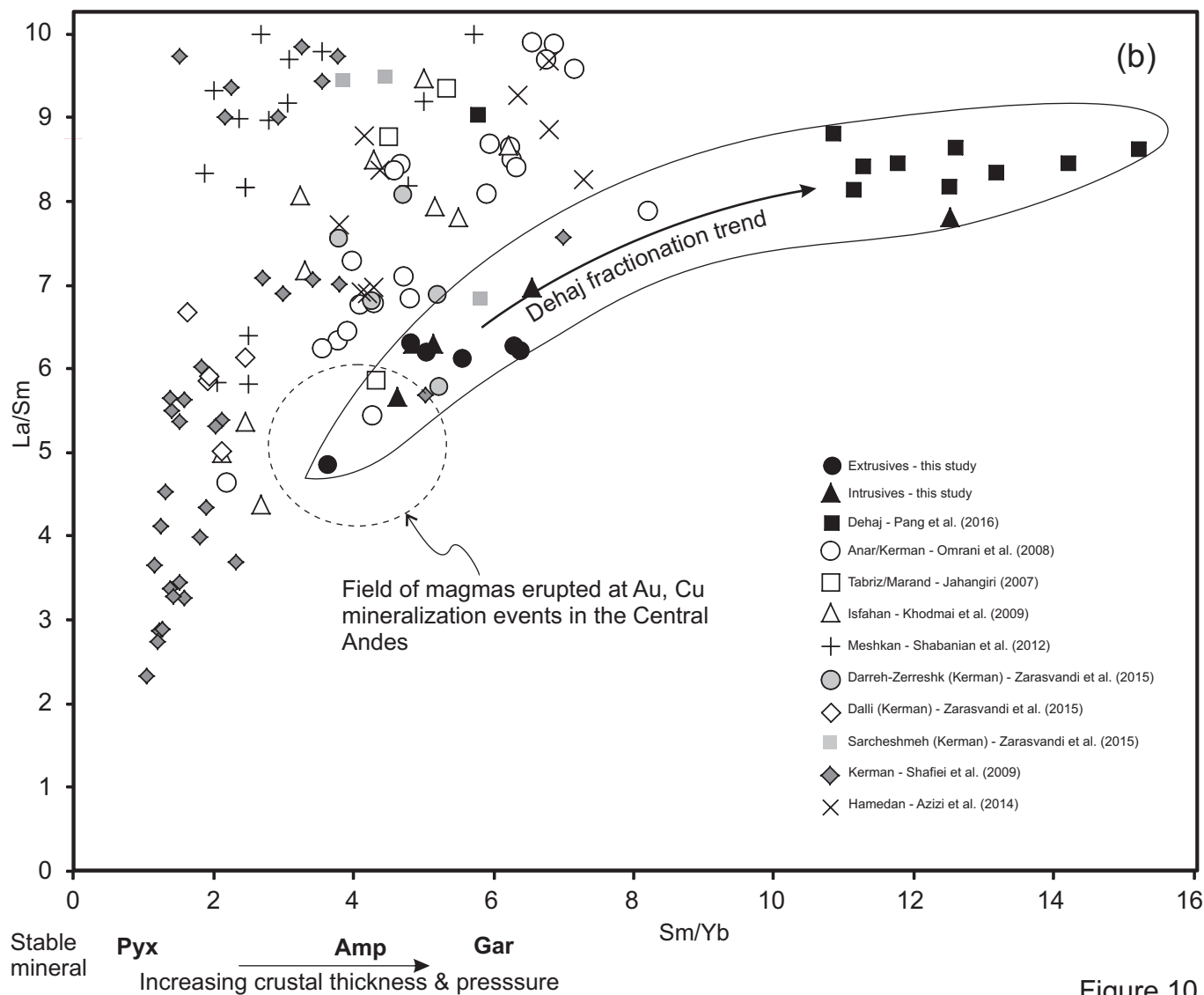
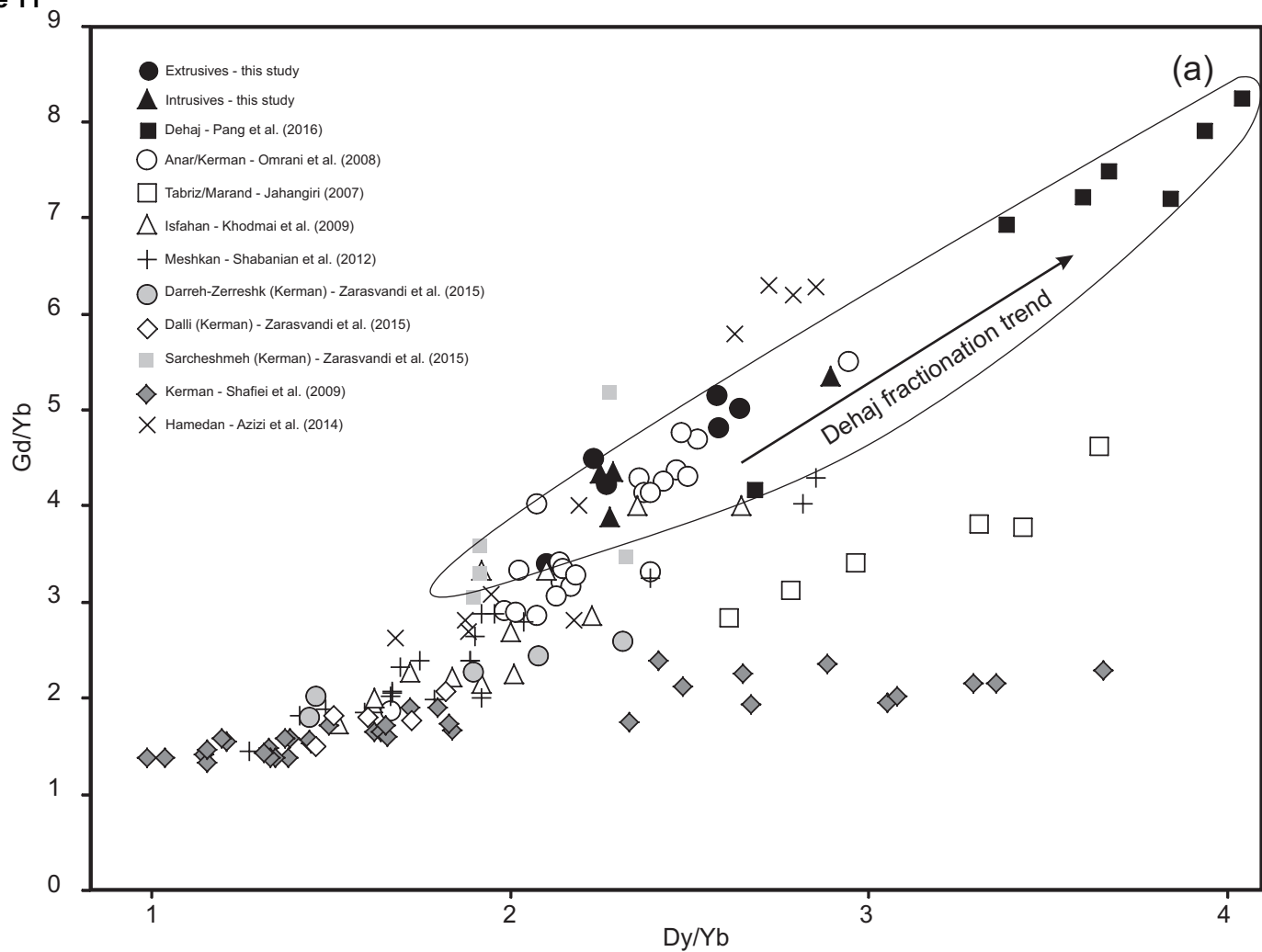


Figure 10

Table 1

Table 1. Kheirkhah et al. Caption in main text.

Sample	DJ1.1	DJ2.1	DJ2.2	DJ6.1	DJ6.2	DJ25.1	DJ25.5	DJ25.7	DJ3.3	DJ4.1	DJ5.1	DJ5.4	DJ7.1
Type	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Flow	Dome	Dome	Dome	Dome	Dome
Lat.	30.721	30.7	30.7	30.752	30.752	30.753	30.753	30.753	30.687	30.67	30.746	30.746	30.84
Long.	54.76	54.755	54.755	54.707	54.707	54.708	54.708	54.708	54.784	54.889	54.791	54.791	54.867
Series	HK	Transit'	Transit'	HK	HK	HK	HK	HK	Transit'	Adakite	Adakite	Adakite	Adakite
<i>Major oxides (wt. %)</i>													
SiO ₂	52.96	60.79	59.95	50.93	54.52	53.20	53.25	54.03	61.77	67.71	71.38	62.16	69.45
TiO ₂	0.97	0.69	0.68	0.82	0.83	0.88	0.82	0.78	0.60	0.43	0.27	0.60	0.32
Al ₂ O ₃	16.23	17.06	16.67	15.28	15.91	15.39	15.48	15.84	16.84	16.54	15.30	17.04	15.59
Fe ₂ O ₃	7.47	5.20	5.23	6.63	6.61	6.53	6.36	6.45	4.69	2.26	1.71	4.63	1.98
MnO	0.13	0.08	0.08	0.13	0.12	0.12	0.12	0.12	0.07	0.02	0.03	0.08	0.04
MgO	6.26	3.14	3.43	4.51	5.87	4.61	4.70	4.75	2.42	0.31	0.43	1.69	0.69
CaO	9.13	6.06	6.52	10.99	8.00	9.36	9.12	9.06	5.47	4.10	2.87	5.67	3.43
Na ₂ O	4.25	4.95	4.89	4.28	4.50	4.29	4.41	4.50	4.88	4.95	4.19	5.53	5.01
K ₂ O	1.62	1.87	1.92	1.89	2.15	2.46	2.27	1.835	2.34	2.46	2.85	1.84	2.77
P ₂ O ₅	0.42	0.34	0.35	0.51	0.51	0.58	0.53	0.47	0.33	0.20	0.09	0.37	0.12
LOI	0.85	0.37	0.56	2.81	0.60	1.91	1.93	1.98	0.62	1.34	1.03	0.50	0.85
Total	100.38	100.56	100.30	99.31	99.71	99.54	99.38	99.96	100.05	100.35	100.41	100.37	100.30
<i>Trace elements (ppm)</i>													
Sr	1029	1071	1129	5095	1718	3104	3673	1681	1001	713.7	721	1532	673.4
Ba	689	600	610	1309	1938	1574	1861	1237	718	599	623	744	608
Y	18.0	12.5	12.7	18.6	18.5	18.9	18.4	18.1	12.5	8.1	4.0	12.6	4.2
Zr	112	139	137	144	157	171	158	147	114	93	57	77	34
Co	31.4	16.7	17.9	30.2	26.8	28.0	29.0	26.1	14.5	4.4	3.0	12.6	4.2
Ni	136.1	54.9	72.3	173.0	148.4	160.6	157.1	135.6	40.1	b.d.	b.d.	5.3	b.d
V	161.0	114.5	116.7	83.7	160.5	109.2	124.2	126.8	96.4	53.9	27.3	88.7	28.3
Cr	323.7	54.8	82.6	260.1	231.1	244.9	233.1	222.0	51.6	7.8	1.8	7.5	3.0
Zn	71.9	68.4	69.2	66.9	76.6	74.1	71.7	69.0	70.8	35.0	37.7	57.3	45.1
Cu	43.5	53.5	76.5	38.2	46.3	73.8	42.1	58.7	33.0	10.5	24.7	3.4	19.7
Sc	15.2	4.4	5.5	12.1	10.3	10.8	10.7	10.4	3.0	b.d.	b.d.	b.d.	b.d.
Rb	33.3	35.0	34.4	28.0	33.8	40.0	37.0	31.0	54.5	59.5	60.6	31.2	62.3
Nb	11.8	6.1	6.4	10.2	9.9	10.3	9.8	9.5	7.4	6.0	2.8	6.8	3.2
La	25.8	36.3	36.9	30.9	45.0	47.8	44.3	42.8	25.6	24.8	21.1	45.1	43.8
Ce	53.6	68.0	68.5	59.3	85.6	93.0	85.4	81.6	49.5	45.2	39.1	86.1	83.8
Pr	6.2	7.4	7.5	6.7	9.3	10.3	9.4	9.0	5.5	4.8	4.1	9.6	9.2
Nd	29.0	34.1	34.7	30.8	42.8	47.9	42.9	41.1	26.0	21.2	17.4	44.1	42.0
Sm	5.3	5.8	5.9	5.0	7.1	7.8	7.2	6.8	4.5	3.6	2.7	7.2	7.0
Eu	1.6	1.6	1.6	1.4	2.0	2.2	2.0	1.9	1.3	1.0	0.8	2.0	2.0
Gd	5.00	4.6	4.8	4.4	6.3	6.8	6.4	6.0	3.8	2.9	2.1	6.1	6.2
Tb	0.6	0.5	0.5	0.5	0.7	0.7	0.7	0.7	0.5	0.3	0.2	0.7	0.7
Dy	3.2	2.4	2.4	2.4	3.5	3.6	3.3	3.3	2.3	1.6	0.9	3.3	3.3
Ho	0.6	0.4	0.4	0.4	0.6	0.6	0.6	0.6	0.4	0.3	0.1	0.6	0.6
Er	1.6	1.0	1.0	1.0	1.6	1.6	1.5	1.6	1.1	0.6	0.3	1.5	1.5
Tm	0.2	0.2	0.2	0.2	0.2	0.3	0.2	0.2	0.2	0.1	b.d.	0.2	0.2
Yb	1.5	0.9	0.9	0.9	1.5	1.5	1.4	1.5	1.0	0.5	0.2	1.4	1.4

Lu	0.2	0.2	0.2	0.1	0.2	0.2	0.2	0.2	0.2	0.1	b.d.	0.2	0.2
Hf	2.93	3.5	3.4	3.8	4.1	4.6	4.3	3.9	3.2	2.5	1.8	2.2	1.0
Ta	0.7	0.3	0.3	0.5	0.5	0.5	0.4	0.5	0.5	0.4	0.3	0.3	0.3
Pb	7.4	10.4	11.1	8.4	13.7	12.7	10.9	10.3	13.3	11.7	16.7	11.4	15.4
Th	4.5	7.2	7.6	10.5	10.6	11.5	10.8	10.9	5.9	5.7	5.3	4.8	4.3
U	1.4	1.9	1.9	2.3	2.7	2.6	2.2	2.4	2.2	2.0	1.4	1.5	0.8

Table 2. Kheirkhah et al. Caption in main text.

	$^{87}\text{Sr}/^{86}\text{Sr}$	$\pm 2\sigma$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\pm 2\sigma$	ϵNd	$\text{SiO}_2 - \text{wt.}\%$	Series
DJ1.1	0.704590	4	0.512832	5	+3.78	53	High-K
DJ4.1	0.704565	8	0.512865	7	+4.43	68	Adakitic
DJ5.1	0.705217	8	0.512942	5	+5.93	71	Adakitic
DJ5.4	0.704624	7	0.512788	5	+2.92	62	Adakitic
DJ6.2	0.705052	10	0.512855	5	+4.23	55	High-K

S

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Iranian plateau

Zagros Suture

Dehaj

Alborz Mountains

Zagros Mountains

Central Iranian
Microcontinent

Moho

Spinel peridotite

Garnet peridotite

Thinner, hotter Eurasian
mantle lithosphere

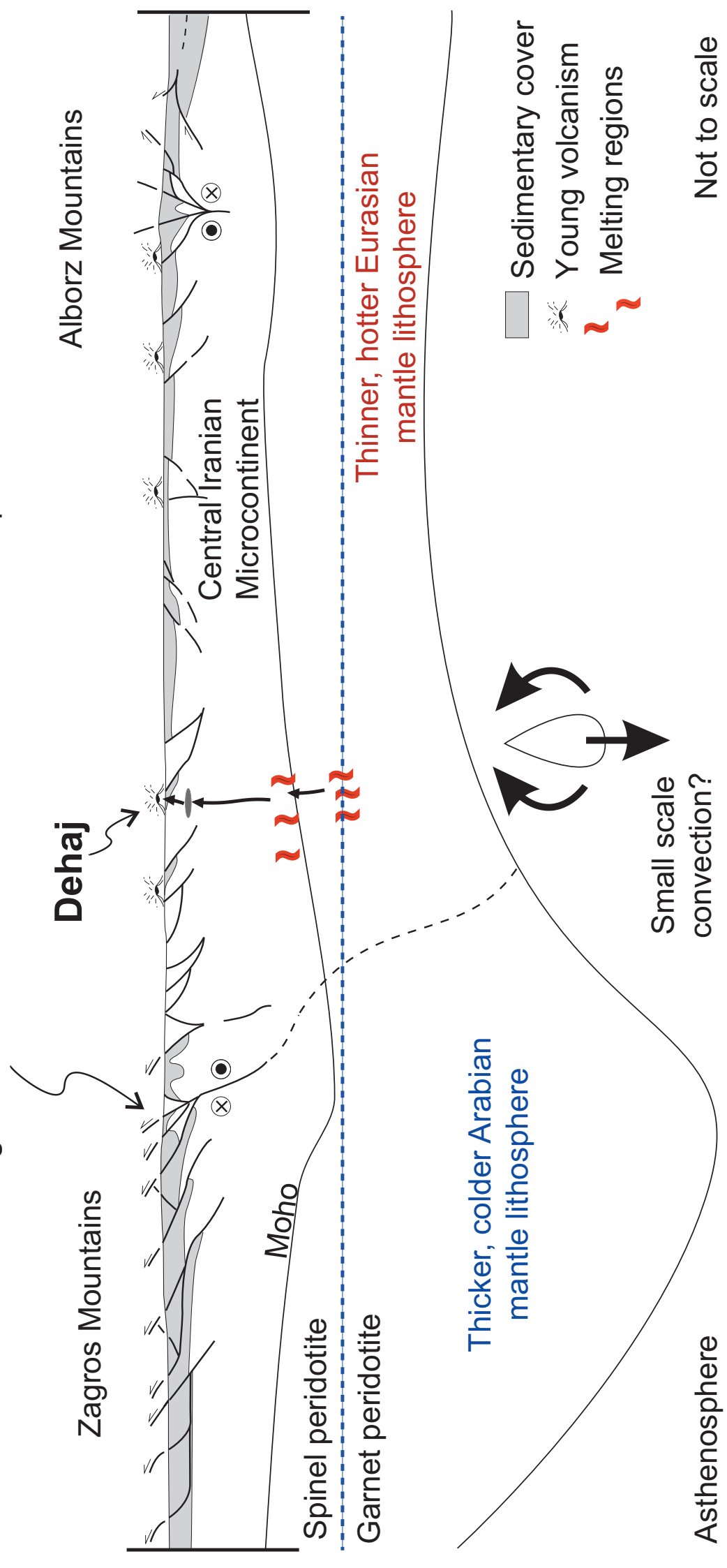
Thicker, colder Arabian
mantle lithosphere

- Sedimentary cover
- Young volcanism
- Melting regions

Small scale
convection?

Asthenosphere

Not to scale



Declaration of interests

☒ The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

☐ The authors declare the following financial interests/personal relationships which may be considered as potential competing interests:

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Monireh Kheirkhah: Conceptualization; data curation; funding acquisition; investigation; project administration; resources; validation; original draft

Iain Neill: Data curation; formal analysis; investigation; methodology; validation; visualization; original draft; review & editing

Mark B. Allen: Conceptualization; Funding acquisition; investigation; project administration; supervision; validation; visualization; original draft; review & editing

Mohammed H Emami: Investigation

Ali Shahraki Ghadimi: Investigation